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CONTENTS

	PAGES
Geology of the Wolf Creek area, St. Elias Range, Yukon Territory, Canada. By Robert P. Sharp	625–650
Origin of sulphides in the nickel deposits of Mount Prospect, Connecticut. By Eugene N. Cameron	651–686
Late Paleozoic faulting in western Newfoundland. By Frederick Betz, Jr.	687–706
Phosphatization at Malpelo Island, Colombia. By Duncan McConnell	707–716

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PAPERS IN PRESS FOR FORTHCOMING ISSUES

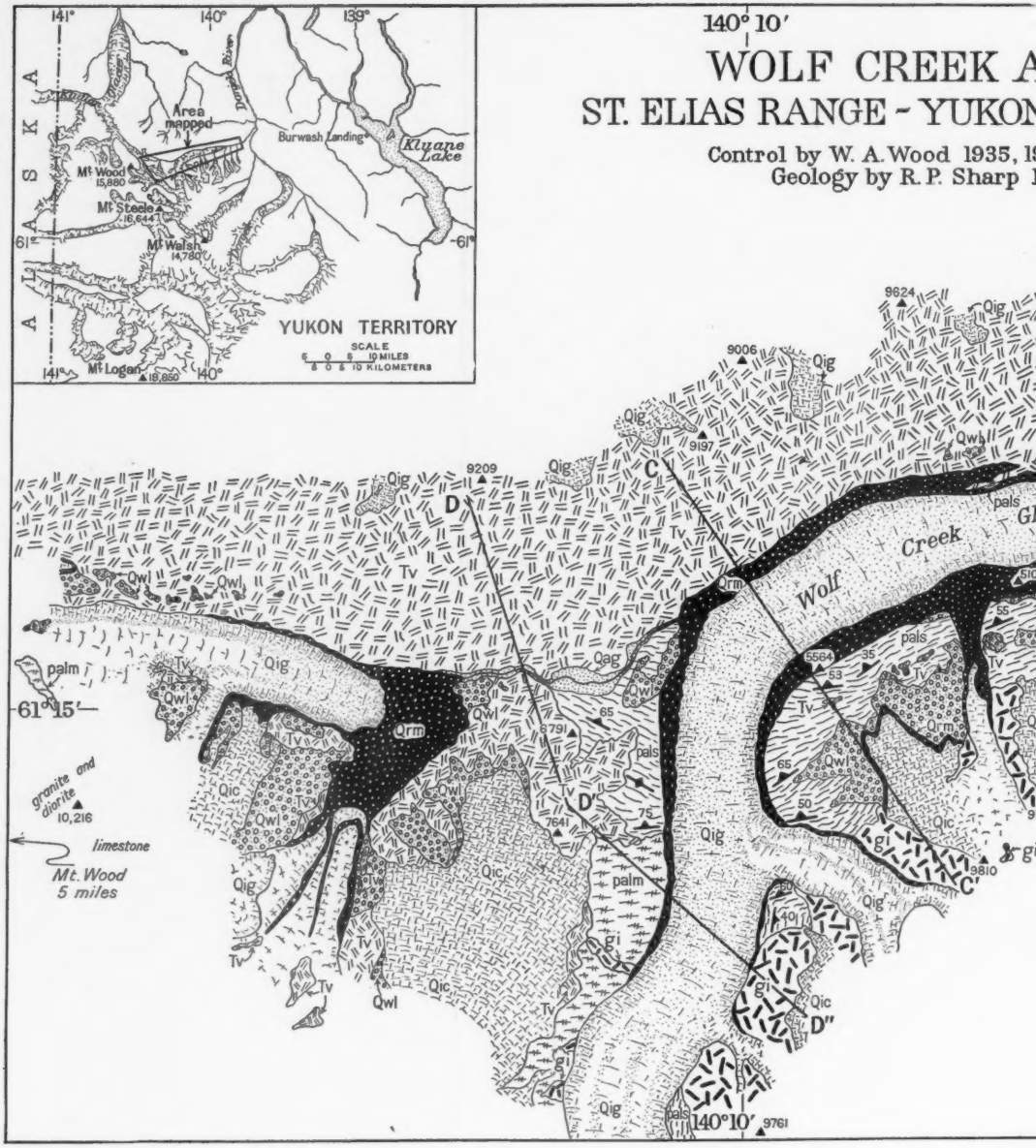
Some Characteristics of the Summit Eruption of Mauna Loa, Hawaii, in 1949. By P. E. Schulz

Transcontinental Gravitational and Magnetic Profile of North America and its Relation to Geologic Structure. By George P. Woollard

Geologic Correlation of Areal Gravitational and Magnetic Studies in New Jersey and Vicinity. By George P. Woollard

Weathering of Igneous Rocks near Hong Kong. By R. W. Brock

Stratigraphy of the Lower Middle Ordovician of Tazewell County, Virginia. By B. N. Cooper and C. E. Prouty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ALASKA
Area mapped
Donjek River
Burwash Landing
Kluane Lake
Mt Wood 15,880
Mt Steele 16,644
Mt Walsh 14,780
Mt Logan 19,850
61°
141°
140°
YUKON TERRITORY
SCALE
10 MILES
10 KILOMETERS
140° 10′
WOLF CREEK A
ST. ELIAS RANGE - YUKON
Control by W. A. Wood 1935, 19
Geology by R. P. Sharp
9824
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Qig
Qwl
Tv
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Qag
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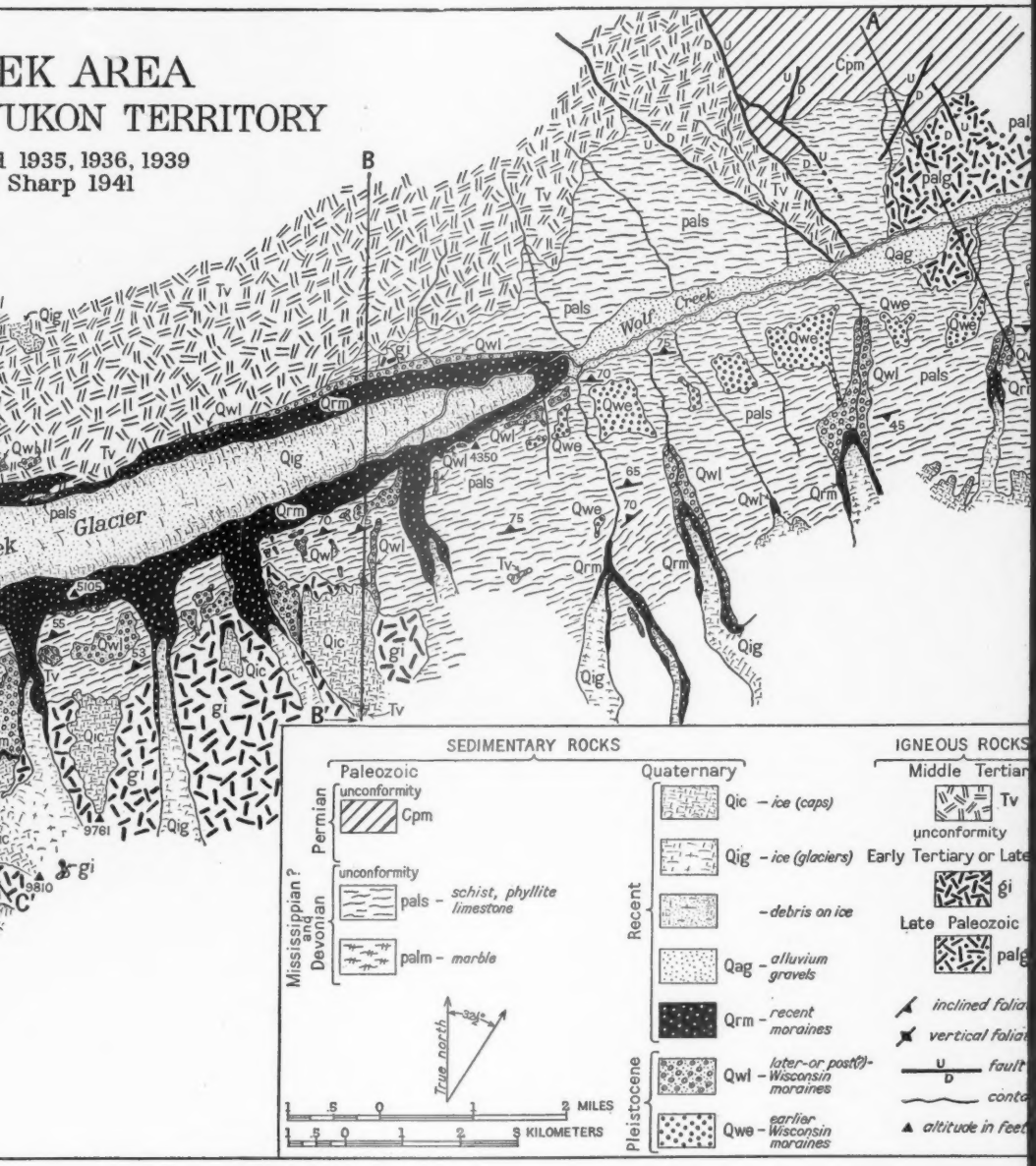
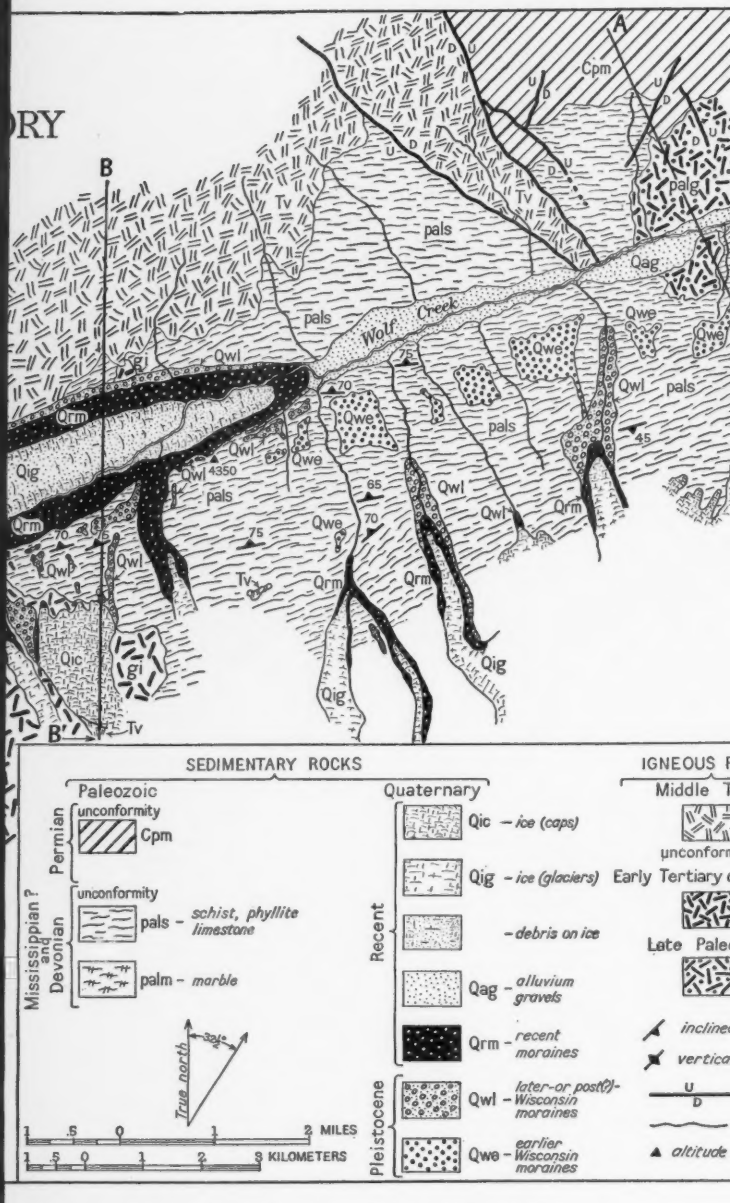
limestone
Mt. Wood
5 miles
140° 10′



GEOLOGICAL MAP AND STRUCTURE SECTION OF WOLF CREEK AREA



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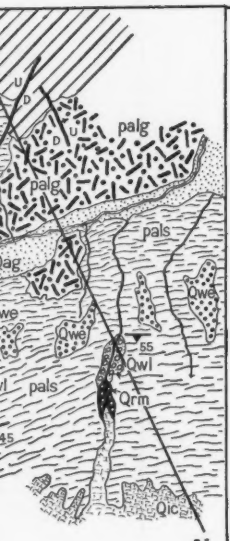


61° 15′

IGNEOUS ROCKS

Middle Tertiary

Tv – *volcanics*

unconformity

Early Tertiary or Late Mesozoic

gi – *granitic intrusives*

Late Paleozoic

palg – *gabbro*

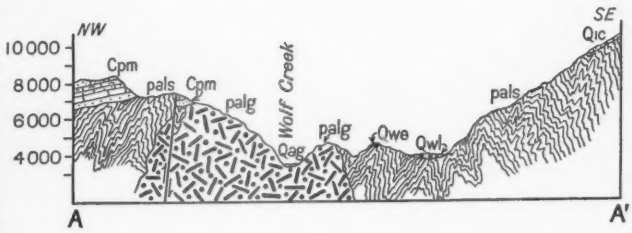
inclined foliation

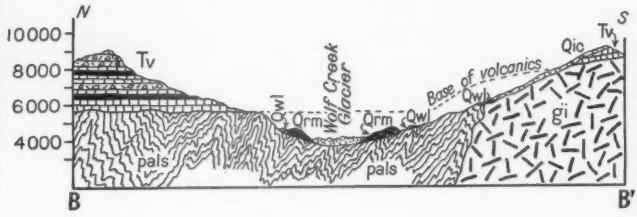
vertical foliation

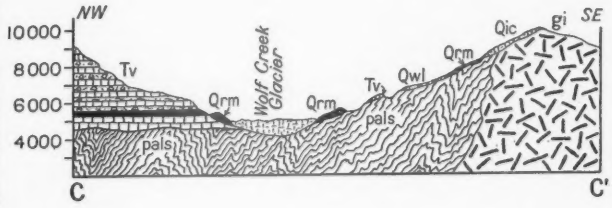
U / D *fault*

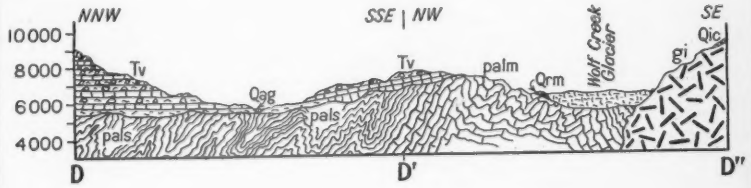
contact

▲ *altitude in feet*









BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 54, PP. 625-650, 5 PLS., 2 FIGS. MAY 1, 1943

GEOLOGY OF THE WOLF CREEK AREA, ST. ELIAS RANGE, YUKON TERRITORY, CANADA

BY ROBERT P. SHARP

CONTENTS

	Page
Abstract	626
Introduction	626
General statement	626
Location and physical features	627
Nature of work	627
Metamorphic and sedimentary rocks	629
Introduction	629
Marble (Devonian)	629
General statement	629
Description	629
Age	629
Undifferentiated metamorphic rocks (Mississippian, Devonian, or both)	630
General statement	630
Description	631
Age and relation to other sections	632
Permian rocks	633
General statement	633
Description	633
Age and relation to other sections	634
Quaternary deposits	635
Igneous rocks	635
General statement	635
Intrusive bodies	635
Pennsylvanian (?) gabbro	635
Late Mesozoic or early Tertiary granitic bodies	636
Tertiary plugs, dikes, and sills	638
Tertiary volcanics	639
General statement	639
Lithology	639
Thickness, arrangement, and structure	641
Origin	641
Pre-volcanic surface	642
Age	643
Structure	644
Summary of geological history	646
References cited	647

ILLUSTRATIONS

Figure		Page
1.	Columnar section	628
2.	Analysis of middle or late Cenozoic stress-strain relations on Wolf Creek	645

Plate Facing page
1. Geological map and structure sections of Wolf Creek area 625
2. Wolf Creek valley 632
3. Rocks of upper Wolf Creek 633
4. Region adjoining Wolf Creek 640
5. Volcanics and structure 641

ABSTRACT

The heretofore unmapped rock sequence on Wolf Creek provides a useful section for correlation of Canadian Yukon and Alaskan formations. The oldest rock, massive Devonian marble at least 1000 feet thick, is overlain by several thousand feet of Devonian-Mississippian slate, phyllite, schist, fissile marble, granulite, and greenstone. An unconformity separates 1500 feet of well-bedded Permian limestone, sandstone, and tuff from the older rocks, and the Permian is overlain discordantly by 4600 feet of well-layered Tertiary volcanics, about half acidic pyroclastics and half andesite and basalt flows.

Pennsylvanian (?) gabbro intrudes the pre-Permian rocks, and large bodies of late Mesozoic or early Tertiary granite and quartz monzonite are exposed. The youngest intrusives are mafic plugs and dikes associated with the Tertiary volcanism. The contact zones of the granitic and monzonitic bodies are pyritized, and a highly fractured silicified area in quartz monzonite contains molybdenite.

The Paleozoic metamorphics are compressed into east-northeast isoclinal folds. The Permian beds are tilted gently northward, and the Tertiary volcanics are largely horizontal. Two sets of high-angle reverse faults trending northwest and northeast near Wolf Creek mouth and an east-northeast monoclinal flexure with upthrow to the south along lower Wolf Creek are major features in the relatively simple structural pattern. The monocline and faults developed in the middle or late Cenozoic and may be the product of adjustments within the St. Elias block during its last major uplift.

INTRODUCTION

GENERAL STATEMENT

During the Wood Yukon Expedition of 1941 (W. R. Wood, 1942a; 1942b; Bates, 1942) the geology of 130 square miles in the northern St. Elias Range was mapped in reconnaissance fashion. This area had been previously visited by a few hunters, prospectors, and two earlier Wood expeditions (W. A. Wood, 1936; F. H. Wood, 1940), but its geology, aside from brief notes by Wood (1936, p. 237), was unreported. The region adjoining Wolf Creek and large areas to the west and south are also blank on the geological map of Yukon Territory (1917). Intermediate location of the Wolf Creek area makes it a valuable link between eastern Alaska and western Canada. This report describes only the bedrock geology, since treatment of the glaciation is reserved for a publication of the American Geographical Society covering activities of all Wood Yukon expeditions.

Every member of the party assisted in the geological work, but special acknowledgment is due Walter A. Wood, who most ably organized and led the expedition. The services of Dr. H. S. Bostock, of the Geological Survey of Canada, who read the manuscript and offered suggestions

based on his wide knowledge of the Yukon, are much appreciated. Dr. J. S. Williams, of the U. S. Geological Survey, kindly studied Permian fossil collections. The University of Illinois Graduate School Research Board aided with a grant defraying expenses of transportation, and the Reserve Officers' Training Corps of the University loaned a small plane table.

LOCATION AND PHYSICAL FEATURES

Wolf Creek, on the northeast flank of the St. Elias Range within the Yukon Territory of Canada, 25 miles east of the Alaskan border, at 61° 15′ N. Lat. and 140° 10′ W. Long. (*see inset* of Plate 1), is one of the larger western tributaries of the Donjek River, a major branch of the White. Whitehorse, the nearest convenient point of entry, lies 180 miles airline east-southeast. Burwash Landing at Kluane Lake, 3 to 4 days pack-train trip—45 miles airline—to the east, served as a final outfitting post.

Peaks of the St. Elias Range, one of the higher and more rugged mountain masses on the continent, culminate in Mt. Logan (19,850). No points on the map (Pl. 1) are over 11,000 feet, although on its western and southern periphery are Mt. Wood (15,880) and Mt. Steele (16,644) (Pl. 2, fig. 1). Relief within the area mapped is about 7000 feet, and complete dissection has produced an extremely rugged topography (Pl. 3, fig. 2). Wolf Creek flows eastward in a broad U-shaped glacial trough (Pl. 2, fig. 1) with a remarkably straight course to the east-northeast in its lower half. The valley makes an almost right-angle bend to the south near its midpoint—herein identified as "the bend."

Wolf Creek valley averages 4000 to 5000 feet deep with floor elevations between 3500 and 6500 feet within the boundaries of Plate 1. Timber line is close to 4500 feet, and fully 90 per cent of the area is free of vegetation large enough to hamper geological work. Orographic snow line lies between 8000 and 9100 feet, and most of the high country is covered by ice and snow (Pl. 3, fig. 1). The only sources of geological information for such areas are a few isolated bare rock ledges and debris carried down by glaciers. A trunk glacier 25 miles long occupies Wolf Creek valley to within 6 miles of its mouth. Tributary canyons indenting the high south wall contain smaller glaciers (Pl. 3, fig. 2), and a larger glacier occupies the major north branch near Mt. Wood (Pl. 1).

NATURE OF WORK

Initially, supplies and equipment were obtained chiefly by parachuted loads from the air (W. A. Wood, 1942a; 1942b), but all subsequent operations were by foot and back pack. Seventy-three days were spent

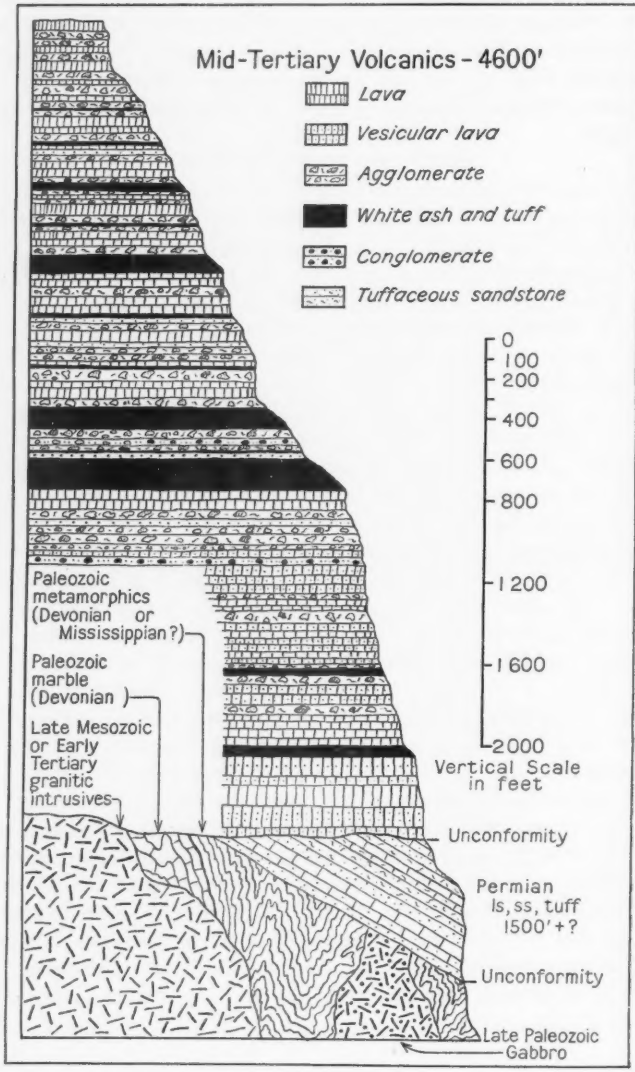


FIGURE 1.—*Columnar section*

on Wolf Creek between June 26 and September 6, 1941. Unfavorable weather handicapped the work only slightly, for the summer of 1941 was one of the driest and clearest reported in Yukon history. Geological mapping was on a 1/50,000 scale controlled by an excellent triangulation system established through theodolitic observations in previous years by Walter Wood. A small U. S. Army plane table and a Brunton compass established locations by resection. Aerial photographs furnished by Wood proved extremely helpful.

METAMORPHIC AND SEDIMENTARY ROCKS

INTRODUCTION

Table 1 summarizes the data on formations mapped along Wolf Creek, and the columnar section (Fig. 1) shows their structural relations.

MARBLE (DEVONIAN)

General statement.—This formation consists of uniform massive gray marble, locally argillaceous and cherty, which is best exposed on the west wall of Wolf Creek above the bend (Pl. 2, fig. 2). The area of exposure is small, but marble may underlie much of the high ice- and snow-covered area to the west. Poorly preserved corals indicate a Devonian age, which is consistent with the stratigraphic succession. Total thickness may be at least 1000 feet and possibly more.

Description.—The marble is uniformly massive, light to dark gray, and of medium grain. It weathers gray, brown, or reddish and contains a few thin argillaceous layers and irregular chert nodules. White coarsely crystalline calcite fills veinlets and replaces bryozoans and corals. Stratification is obscured by fracturing, metamorphism, and weathering. The topographic habit of the marble is characterized by steep, rough surfaces with numerous spires and sharp ridges.

Age.—A species of *Cladopora* and one other distorted coral with Devonian characteristics are among the poorly preserved fossils in the marble. Dr. W. H. Easton of the Illinois Geological Survey identifies these corals as definitely pre-Carboniferous and probably not older than Devonian.

The massive marble is overlain conformably by a considerable thickness of schist, phyllite, and thin marble beds unless the entire section on Wolf Creek is inverted which seems unlikely. The schist and associated rocks unconformably underlie Permian beds and are considered Mississippian or Devonian, or both. The probable Devonian age of the fossils is thus consistent with the stratigraphic succession. Massive Middle Devonian limestone in various parts of Alaska (Brooks and

TABLE 1.—*Summary of formations*

Quaternary	Recent gravels (Qag)	Few hundred feet thick at most, largely bouldery outwash gravels of modern glaciers
	Recent moraines (Qrm)	100 to 200 feet, fresh, bouldery till of present glaciers and modern advance 100 to 200 years ago
	Later or post-Wisconsin moraines (Qwl)	100 feet, bouldery till, slightly weathered, covered with vegetation, exact age unknown
	Earlier Wisconsin moraines (Qwe)	100 feet, bouldery till, noticeably weathered, subdued topography, 500 to 1500 feet above present streams, older than preceding later or post-Wisconsin moraines, but still probably late Wisconsin
Tertiary	Volcanics (Tv)	4600 feet, well-layered basalt, andesite, trachyte, quartz latite, obsidian, ash, tuff, agglomerate, breccia, all of local origin, volume of flows and pyroclastics about equal, mafic dikes, sills, plugs associated
Early Tertiary or Late Mesozoic	Granitic intrusives (gi)	Large batholithic body of porphyritic biotite granite and hornblende-biotite quartz monzonite, locally mineralized and containing molybdenite
Permian	Sedimentary rocks (Cpm)	1500 feet, well-stratified limestone, limy sandstone, sandstone, grit, chert, sandy tuff, and water-laid tuff, containing Permian fossils
Pennsylvanian (?)	Gabbro (palg)	Stocklike body of medium- to coarse-grained augite-hornblende gabbro, intruding pre-Permian rocks
Mississippian, Devonian, or both	Undifferentiated metamorphic rocks (pals)	Several thousand feet of slate, phyllite, schist, fissile marble, quartzite, greenstone, and granulite, originally sedimentary and volcanic, closely and complexly folded
Devonian	Marble (palm)	1000 feet or more, massive, uniform, gray marble, slightly cherty and argillaceous, sparse corals

Kindle, 1908, p. 277-279; Smith, 1939, p. 18-25; Capps, 1940, p. 103-104) may be in part equivalent to the marble of Wolf Creek.

UNDIFFERENTIATED METAMORPHIC ROCKS (MISSISSIPPIAN, DEVONIAN, OR BOTH)

General statement.—This formation consists of slate, phyllite, schist, marble, quartzite, greenstone,[1] and granulite[2] produced by a low-grade

[1] Greenstone is used here for a relatively massive rock, containing enough chlorite and epidote to give it a greenish color, which was produced by low-grade metamorphism of fine-grained mafic igneous material.

[2] Granulite is used in the sense defined by Holmes (1928, p. 111): "A granulose metamorphic rock composed of even-grained interlocking granular minerals."

regional and somewhat more intense contact metamorphism of sedimentary and volcanic beds. It outcrops over a large part of the area mapped (Pl. 1) and composes much of the south wall of Wolf Creek below the bend. Most of the rocks are schistose and so complexly folded that measurement of thickness is impossible. On the basis of stratigraphic relations this unit is Devonian or Mississippian, or possibly both.

Description.—Marble and phyllite dominate. The marble is gray to nearly black, finely crystalline, fissile, schistose, and crenulated. It weathers gray to brown. Impure calcareous beds contain dolomite, sand, clay, and carbonaceous material. Veinlets of coarse-grained white calcite are numerous in fractured strata, and small pods of hydrothermal quartz are locally abundant. Black, paper-thin phyllite forms partings between layers of marble, and phyllite beds in groups 200 to 300 feet thick alternate with similar groups of marble strata. Most phyllite is shiny gray to nearly black, brownish weathering, and thin-bedded. Paper-thin beds are rich in clay, and platy strata contain some sand and calcite. Small-scale sharply asymmetrical crenulations, less than 0.5 mm. high, give many phyllites a prominent linear marking on schistosity planes, and cross crenulations locally produce a reticulated pattern. Slates are not abundant, although much of the phyllite is slightly recrystallized and resembles slate. Scattered beds of gray, brown-weathering, fine, argillaceous quartzite were noted.

Near intrusives phyllite and marble gradually give way to schist and granulite. Quartz-biotite, quartz-sericite, and sericite-chlorite schists of medium grain are common. The various granulites consist of interlocking granules of calcite, quartz, diopside, hornblende, biotite, chlorite, zoisite, sericite, plagioclase, and traces of pyrite. In some beds a distinct compositional interlayering of biotite and quartz with pyroxene and calcite is apparent.

Greenish calcareous sericite-chlorite schists and greenstones crop out on lower Wolf Creek. The greenstones consist of chlorite, epidote, hornblende, plagioclase (near albite), sericite, calcite, sphene, and traces of pyrite in interlocking granules without well-developed foliation. Veinlets and pods of coarse-grained hydrothermal quartz and ankerite(?) are abundant in these rocks.

The rocks composing this formation were developed by low-grade regional metamorphism of shale, calcareous sandy shale, limestone, impure sandy and dolomitic limestone, and argillaceous sandstone. In lower Wolf Creek, mafic volcanics, presumably flows or tuff, and possibly tuffaceous limestone were included. Granitic intrusions locally produced more intense metamorphism. These rocks are so tightly and complexly

folded that the thickness cannot be determined accurately. One to several thousand feet appears a not improbable figure.

Age and relation to other sections.—The only direct field evidence on the age of this unit is its conformable position above marble containing Devonian fossils and its unconformable attitude beneath Permian beds.

Metamorphic rocks of the pre-Mesozoic complex east and north of Wolf Creek are usually assigned to the Yukon group of pre-Cambrian or Paleozoic age (Lees, 1936, p. 8-11; Bostock, 1936, p. 14-19). The Mount Stevens group (Cairnes, 1912, p. 40-46; Cockfield and Bell, 1926, p. 8-10) in the Wheaton and Whitehorse districts and previously established formations of undetermined age, such as the Kluane schist (McConnell, 1905, p. 4A-5A; Cairnes, 1915b, p. 14-15), are probably Yukon equivalents. The metamorphics of Wolf Creek are somewhat similar to the Yukon group, but Bostock (personal communication) thinks the descriptions above indicate less marked regional metamorphism than shown by that group along Yukon River. This need not be a fatal objection to correlation, but there is yet no proof that the Wolf Creek rocks should be included in the Yukon group.

In Alaska deposits of every Paleozoic period are known (Smith, 1939), and many of the systems contain rocks similar to the Wolf Creek metamorphics. However, pre-Devonian rocks may be tentatively eliminated, for the Wolf Creek metamorphics overlie Devonian marble, and comparable pre-Devonian rocks are scarce in the vicinity of Wolf Creek. In contrast, Middle Devonian rocks are widespread in southeastern Alaska (Moffit, 1938b, p. 18), and the Mississippian is well represented throughout the Territory (Mertie, 1930b, p. 84; Smith, 1939, p. 25). Both systems contain rocks similar to the Wolf Creek metamorphics. They also include metavolcanics (Capps, 1916, p. 31; Mertie, 1935, p. 298, 300; Moffit, 1938a, p. 23) like those on Wolf Creek and are cut by similar gabbroic intrusives (Mertie, 1937, p. 204-206; Moffit, 1941, p. 125). Since there is little reason for excluding either the Devonian or Mississippian of Alaska as possible correlatives, it is best to assume that the Wolf Creek rocks may be either Mississippian or Devonian or possibly in part both. This is consistent with the Devonian age of the underlying marble, the lack of definite break between Devonian and Lower Carboniferous beds (Brooks and Kindle, 1908, p. 291, 312), and the unconformable position of the overlying Permian.

Similar rocks along the Donjek River just to the east contain Carboniferous fossils (McConnell, 1906, p. 22). This relation and the abundance of Mississippian rocks on upper Chitina River 35 miles west (Moffit, 1938a, Pl. 2) might suggest that the Wolf Creek metamorphics are largely Mis-

FIGURE 1. WOLF CREEK VALLEY, MT. STEELE(S), MT. WOOD(W)
Donjek flood plain and Wolf Creek fan behind foreground ridge, stagnant lower Wolf Creek Glacier in center. Looking west-southwest. (Aerial photograph by W. A. Wood, Sept. 9, 1941.)

FIGURE 2. ROCKS OF UPPER WOLF CREEK
Foreground ridge composed of dark horizontal Tertiary volcanics on Devonian marble. Peaks beyond Wolf Creek Glacier largely granitic. Looking east. (Aerial photograph by W. A. Wood, Sept. 9, 1941.)

FIGURE 1. WOLF CREEK GLACIER ABOVE BEND
Chiefly quartz monzonite and some Devonian-Mississippian metamorphics to left of glacier. Peak in right center is Mt. Walsh. Looking southeast. (Aerial photograph by W. A. Wood, Sept. 9, 1941.)

FIGURE 2. EAST WALL OF WOLF CREEK NEAR BEND
Continues northward from Figure 1. Rocks largely granite and quartz monzonite. Looking east-southeast. (Aerial photograph by W. A. Wood, Sept. 9, 1941.)

ROCKS OF UPPER WOLF CREEK

sissippian. Lack of definite correlation between the fossiliferous Donjek beds and the Wolf Creek rocks and the possibility that McConnell's fossils may have come from strata now known to be Permian lend uncertainty to this dating. The Wolf Creek rocks are probably not younger than Mississippian, for they are unconformable beneath Permian strata, and the Pennsylvanian is poorly represented in Alaska. The only beds definitely known to be Pennsylvanian are terrestrial (J. S. Williams, 1941, p. 1978).

PERMIAN ROCKS

General statement.—Permian strata are exposed on the north wall of lowermost Wolf Creek (Pl. 1). The following data were obtained from distant observations and study of float, for access to the Permian outcrops was prevented by Wolf Creek, which in summer is unfordable by foot. Abundant float and excellent exposures permit reasonably reliable descriptions.

About 1500 feet of well-stratified limestone, limy sandstone, grit, chert, sandy tuff, and water-laid tuff are exposed (Pl. 5, fig. 2). Limestone, sandstone, and tuff are about equally abundant, and Permian fossils are plentiful.

Description.—The 1500 feet (estimated) of Permian beds appear from a distance to be as follows:

Permian strata (descending order)	Feet
7. Massive black beds	50
6. Uniform, well-bedded, gray-weathering strata	500
5. Interlayered yellowish-brown, white, gray, and black beds, constituting the most distinctive group within the Permian sequence	100
4. Dark, massive beds	50
3. Dark-weathering, massive beds with some light-colored strata	250
2. Dark-weathering, well-bedded strata with two prominent white layers, 20 to 40 feet thick	250
1. Brown-weathering, well-bedded strata	300
Total	1500

The calcareous strata are light-gray, fine-grained, soft, argillaceous limestone; black, dense, brittle limestone with irregular fracture; and various shades of gray, medium- to fine-grained arenaceous and tuffaceous limestone. Most of these beds weather light brown or gray. The large fragments of chert suggest that it occurs as beds rather than solely as nodules. Gray, green, white, and brown dense chert layers are interbanded on a ½- to 1-inch scale. Arenaceous strata consist of light-gray to brownish-gray, fine- to medium- and even-grained quartz sandstone some of which is tuffaceous and some calcareous. Nearly all the sandstone is distinguished by a vivid warm-brown weathering color. Tuff beds are gray, fine-grained, weather brown, and contain considerable quartz and feldspar. The tuff is well cemented and lacks the lami-

nation noted in calcareous strata, although it appears to be well bedded in outcrop and is presumably water laid. The grits are brownish to greenish gray and consist of subangular to subrounded fragments of quartz, tuff, limestone, chert, and unidentified materials up to ⅜ inch in diameter. The matrix is variously sandy, calcareous, or tuffaceous. The fossiliferous beds are dark-gray, arenaceous limestone and light- to brownish-gray, limy sandstone.

The northern St. Elias Range contains a greater thickness of Permian strata than the 1500 feet exposed on Wolf Creek, for these beds extend northward in a homoclinal section, and much greater thicknesses have been reported in Alaska (Capps, 1916, p. 39; Ross, 1933, p. 295; Moffit, 1938b, p. 19).

The Permian beds of Wolf Creek have a gentle northward tilt and are cut by a number of high-angle oblique faults. They are unconformable on Paleozoic gabbro and metamorphics at the mouth of the canyon and discordantly overlain by Tertiary volcanics. The pre-Permian erosion surface is smooth throughout the short distance exposed. Unconformities between Permian and Mississippian or Devonian rocks are reported in Alaska (Moffit, 1938a, p. 20; Ross, 1933, p. 294).

Age and relation to other sections.—Fossils collected in float have been studied by Dr. J. S. Williams, who kindly furnishes the following statement.

"The collection consists of species of the brachiopod genera: *Chonetes, Productus, Marginifera, Spirifer* (*Neospirifer*), *Punctospirifer,* and *Rhynchopora.* Two species of bryozoans are also present. The brachiopod species are, for the most part, as yet unpublished, but they are well known in Alaskan collections of late Paleozoic age. The faunule is typical of the Alaskan faunules at present generally referred to the Permian period."

Permian strata are reported in many Alaskan districts within a 300-mile radius of Wolf Creek other than those already referred to (Capps, 1940, p. 107; Moffit and Knopf, 1910, p. 17-22; Moffit, 1933, p. 144-148; 1941, p. 129-132; Mertie, 1930b, p. 121-130; 1937, p. 146-153), and beds of possible Permian age are known in the Canadian Yukon (Cairnes, 1915a, p. 71, 82; Bostock and Lees, 1938, p. 11). Brooks' (1900, p. 359-360) Nutzotin series and Carboniferous strata near Whitehorse (Cockfield and Bell, 1926, p. 12-14) may also contain deposits of this period. Strong lithologic similarities between the Wolf Creek rocks and Permian strata in Alaska suggest at least partial equivalence. The middle division of the Alaskan Permian, consisting of intermixed sedimentary and volcanic beds, seems to be represented here rather than either the upper or lower divisions which are predominantly volcanic (Moffit, 1930, p. 147; Ross, 1933, p. 295-298). This may indicate a Permian overlap to the east or southeast.

QUATERNARY DEPOSITS

Thorough treatment of Quaternary materials requires a discussion of glacial history beyond the scope of this article. Deposits of two Wisconsin substages are mapped and designated for convenience as earlier and later Wisconsin. The older moraines resemble those of the Iowan-Wisconsin advance in the mountains of western United States (Sharp, 1941, p. 1932-1933), but other considerations suggest that they may not be older than late Wisconsin.[3] The younger substage is latest or possibly post-Wisconsin and includes moraines of two separate but closely related phases. Recent deposits of bouldery till and outwash gravels are related to existing glaciers and to a marked advance culminating 100 to 200 years ago. This dating is based in part on ecological data and parallels the established history of coastal Alaskan glaciers, many of which drain from the same snow fields.

The Recent white volcanic ash reported throughout the Yukon (Dawson, 1888, p. 45B; Hayes, 1892, p. 148; Cairnes, 1912, p. 73-74) is absent from Wolf Creek. According to Capps (1915, p. 61) the source was about 40 miles northwest, and his isopach map indicates a possible 2 to 3 feet of ash in this vicinity. Bostock (personal communication) suggests from his observations that wind direction during eruption may not have favored deposition on Wolf Creek. Rugged local terrain, exceptionally active erosion, and the possibility that the ash fell on snow which subsequently melted and washed it away are supplementary or alternate factors. This 1400-year-old ash (Capps, 1916, p. 83) is one of the best chronological markers in the region.

IGNEOUS ROCKS

GENERAL STATEMENT

Igneous rocks on Wolf Creek are classified into four age groups, the oldest being the metamorphosed Devonian-Mississippian mafic flows or tuffs already described. A Pennsylvanian(?) gabbroic stock intrudes the Paleozoic metamorphics, and considerable late Mesozoic or early Tertiary granite and quartz monzonite are exposed. Tertiary lavas and pyroclastics cover a large area north of Wolf Creek and with associated dikes, sills, and plugs constitute the youngest igneous group.

INTRUSIVE BODIES

Pennsylvanian(?) gabbro.—A stocklike body of gabbro composes the north wall of Wolf Creek at its mouth (Pl. 1). The rock is a dark-green altered augite-hornblende gabbro of xenomorphic granular texture and medium grain size, containing finer diabasic phases and extremely coarse

[3] Subsequent studies tend to confirm the late Wisconsin age of the older moraines.

pegmatitic segregations. Secondary serpentine, chlorite, epidote, and calcite are abundant, and some talc is seen in hand specimen and thin section.

Exact dimensions of the gabbro body cannot be determined, but it is discordant and seemingly small, so it may provisionally be classed as a stock. The gabbro intrudes the Devonian-Mississippian metamorphics and is unconformably overlain by Permian beds. It is certainly late Paleozoic and probably Pennsylvanian, for absence of marine Pennsylvanian beds in Alaska (J. S. Williams, 1941, p. 1978) suggests widespread diastrophism and probably igneous intrusion during that period. Similar basic and ultrabasic Paleozoic bodies are known in Alaska (Mertie, 1937, p. 199-200, 203-210; Smith, 1939, p. 88), although not all are contemporaneous.

Late Mesozoic or early Tertiary granitic bodies.—The south wall of Wolf Creek westward from its midpoint is composed largely of granite and monzonite (Pl. 1). Aerial photographs suggest a considerable southward extension of the granitic rocks (Pl. 2, fig. 2; Pl. 3, fig. 2), and debris on the glaciers shows that many of the high snow- and ice-covered peaks to the southwest are composed of similar material. A body of batholithic proportions is indicated.

The chief rocks are a pink porphyritic biotite granite between middle Wolf Creek and the bend and varieties of white to gray hornblende-biotite-quartz monzonite farther south. These rocks are of medium grain and hypautomorphic texture. Myrmekite, perthite, and zoned plagioclase are common. The porphyritic granite has pink orthoclase phenocrysts 1½ inches long, and a melanocratic phase of the quartz monzonite contains considerable sphene. Gneissic structure and inclusions are prominent in the granitic rocks only near intrusive contacts. Small aplite and pegmatite dikes, some with black acicular tourmaline, were noted, and three plugs of dark biotite-hornblende tonalite containing considerable sodic oligoclase intrude the Devonian marble on the west wall of upper Wolf Creek. One plug is truncated by the erosion surface below the Tertiary volcanics, and since the tonalite is peripheral to the granitic rocks it is assigned to the same group. Time did not permit separate mapping of the various lithologies within this group, so they are not differentiated on Plate 1.

The porphyritic biotite granite is highly silicified, pyritized, and contains a few flakes of molybdenite and chalcopyrite in a 200- to 300-foot border zone north and east of peak 9761 (Pl. 1). A large medial moraine on Wolf Creek Glacier consists of highly silicified quartz monzonite containing chalcopyrite, pyrite, specular hematite, and considerable mo-

lybdenite in small veinlets and flakes. The abundance of mineralized debris in the moraine indicates a large exposure farther up the valley outside the map area (Pl. 1). Molybdenum is reported in 41 localities in Alaska, some not far west of Wolf Creek (Smith, 1942, p. 161-164).

The granitic rocks intrude the Devonian-Mississippian metamorphics, but their relationship to the Permian beds is not known. They are unconformably overlain by Tertiary volcanics, and intrusion probably occurred before deposition of the lignite-bearing late Eocene beds (Cairnes, 1915b, p. 15) a short distance east of Donjek River as indicated by the lack of metamorphism and marked deformation in the latter. On a stratigraphic basis the Wolf Creek granitic rocks may be tentatively dated as post-Devonian or Mississippian and pre-late Eocene. The lack of Paleozoic intrusives of this type in the Alaskan Yukon (Smith, 1939, p. 88-89) and other relations indicate that they are probably post-Paleozoic.

In the Alaskan Yukon, post-Paleozoic intrusives are broadly grouped as Mesozoic and Cenozoic (Smith, 1939, p. 88), but Mertie (1930a, p. 106) and Moffit (1938a, p. 106-107) show that a three-fold subdivision is feasible. On this latter basis the following classification is formulated: (1) Coast Range batholith rocks, late Mid-Jurassic to early Lower Cretaceous; (2) Laramide rocks, late Upper Cretaceous to early Eocene; and (3) mid-Tertiary intrusives.

The Coast Range rocks form large batholiths and satellitic bodies predominantly of granodiorite or quartz diorite, although granite, monzonite, syenite, and acidic and basic differentiates are known. Smith (1939, p. 88) states that quartz monzonite is rare in the Coast Range bodies of the Alaskan Yukon, but this generalization may not hold elsewhere (Cairnes, 1912, p. 52; Bostock, 1936, p. 34). The Laramide bodies are also of batholithic proportions and largely granite and monzonite, with a variety of other rocks reported locally. The Laramide and Coast Range bodies are alike in size but differ in average composition. Somewhat similar relations are recognized in western United States (Lindgren, 1915, p. 254-261).

The age of the Alaskan bodies herein termed Laramide has not been accurately determined, and failure in the past to distinguish them either from Coast Range or mid-Tertiary intrusives leads to confusion. Capps (1933, p. 286) reports intrusion of granite during the Eocene in the Mount McKinley region, and Mertie (1938, p. 86) favors an early Tertiary age for seemingly correlative intrusives farther west. The bodies of granite, monzonite, and diorite near Broad Pass (Moffit and Pogue, 1915, p. 58-59) previously dated as post-Eocene because they intrude the Cantwell formation need be only late or post-Cretaceous, as the

Cantwell is now known to be Upper Cretaceous and not Eocene as formerly supposed (Smith, 1939, p. 63; Capps, 1940, p. 118). Large intrusives of late Cretaceous or early Tertiary age have also been reported in Suslota Pass (Moffit, 1933, p. 158), possibly in the upper Chitina River (Moffit, 1938a, p. 106-107), and in the Yukon Plateau to the east of Wolf Creek (Bostock and Lees, 1938, p. 18-19). The existence of a number of large intrusives approximately of Laramide age (late Upper Cretaceous to early Eocene) appears to be established.

Mid-Tertiary stocks, laccoliths, dikes, and sills of granite, quartz monzonite, and monzonite are exposed in the Ruby-Kuskokwim (Mertie, 1924, p. 69-71), Yukon-Tanana (Mertie, 1937, p. 219-226), and possibly the Nushagak (Mertie, 1938, p. 87) districts. Many are related to cinnabar mineralization, and some still possess peripheral hot springs.

The size, composition, lack of cinnabar mineralization or hot springs, and inferred antecedence to the coal-bearing Eocene beds east of the Donjek indicate that the Wolf Creek intrusives are not mid-Tertiary. Furthermore, the granitic intrusives of east-central Alaska and most molybdenite-bearing bodies of that region are now considered Mesozoic (Smith, 1942, p. 205). The predominance of granite and quartz monzonite in the Wolf Creek bodies favors correlation with the Laramide group, but Laramide rocks are scarce in near-by areas in contrast to the abundance of Coast Range rocks. It seems prudent for the present to recognize that the Wolf Creek bodies may be either Laramide or Coast Range (late Middle Jurassic to early Eocene) with intrusion during the Mesozoic considered most likely.

Tertiary plugs, dikes, and sills.—Coarse-grained extensively altered sodic monzodiorite containing anorthoclase and distinctive titanium-rich augite composes small pluglike bodies at the base of the Tertiary volcanic rocks in upper Wolf Creek. Sills of diabase within the volcanics near the same locality and a diabase plug intruding schists near point 5564 (Pl. 1) south of middle Wolf Creek contain the same titanium-rich augite. This suggests a common source for the monzodiorite and diabase. A number of mafic dikes 2 to 20 feet wide cut the volcanics and adjacent Paleozoic rocks. Those examined in thin section are augite andesite or diabase of fine to coarse diabasic texture. These intrusives are of essentially the same composition as the Tertiary flows and presumably of the same age.

A rhyolite porphyry dike in Paleozoic schist near peak 6791 may possibly be related to similar rocks of Cenozoic age in other areas (Cairnes, 1915a, p. 102-103; Cockfield, 1921, p. 33-35; Mertie, 1937, p. 216-219, 226-228).

TERTIARY VOLCANICS

General statement.—Well-layered nearly horizontal volcanic rocks at least 4600 feet thick compose much of Wolf Creek's north wall (Pl. 5, fig. 1). Small outliers are preserved south of the canyon, and volcanics cover a large area west of Wolf Creek bend (Pl. 1). Pyroclastics, largely tuff and agglomerate, compose at least half this sequence, the remainder being basalt and andesite flows. Quartz latite, trachyte, and obsidian are minor constituents. The volcanics are Tertiary, of local origin, and widely distributed in this part of the St. Elias Range (Pl. 4, fig. 1) (Brooks, 1900, p. 362-363; McConnell, 1905, p. 8A).

Lithology.—Most of the flows are augite andesite and olivine-augite basalt with andesite slightly predominant. A thick flow of sodic quartz latite and a group of trachyte or possibly sodic latite flows resemble soda-rich rocks reported in corresponding volcanics of the Nabesna-White River District, Alaska (Moffit and Knopf, 1910, p. 35). Except for obsidian, all the lavas are holocrystalline with fine hypautomorphic granular, pilotaxitic, or trachytic textures most abundant. Porphyritic types are slightly more abundant than nonporphyries. Common phenocrysts in andesite are augite, zoned labradorite, and less abundantly biotite. In addition, olivine and more rarely basaltic hornblende are present in the basalt flows. The quartz latite contains large crystals of quartz, anorthoclase, andesine, and aegirine-augite. A coarse-grained amygdaloidal basalt near the base of the section is distinguished by its tabular crystals of amber glassy labradorite ¾ inch long. Similar phenocrysts are reported from lavas in neighboring parts of Alaska (Moffit and Knopf, 1910, p. 36; Moffit, 1938a, p. 94). Pyrrhotite is the major ore accessory in the lavas, and cristobalite fills small cavities in the quartz latite and one andesite flow.

Some flows are moderately vesicular, and others are scoriaceous with horizontally flattened vesicles. Roughly spherical amygdules, up to 1½ inches in diameter, are filled with quartz, chalcedony, calcite, and aragonite. However, most flows are neither vesicular nor amygdaloidal. Columnar jointing is locally well developed in a few flows near the base of the sequence.

Obsidian showing flow lines and crude layering outcrops in cliffs 200 feet high on the flanks of peak 6791 west of Wolf Creek bend and along the river banks to the north (Pl. 1). This rock is chiefly a black vitreous to pitchy glass containing scattered crystals of potash feldspar and an irregular intermixture of 5 per cent reddish-brown glass. Under the microscope the red material appears to be small braided layers, stringers, and spicules of hematite which represent the oxidized black particles of

similar size and arrangement in the black glass. The indices of the red (1.495) and the black (1.498) glass show that they are both obsidian, with about 73 per cent silica (George, 1924, p. 365). Long splinters of gray obsidian associated with tuff near the middle of the volcanic section farther east contain crystals of potash feldspar and have slightly higher index (1.507).

The pyroclastics range from fine ash and tuff to coarse agglomerate and breccia. The fine deposits are white, gray, or brick red and usually well bedded, although some are massive. Ash beds are soft and friable, but the tuff layers are indurated, and some near the base of the sequence in the west are hard, platy, and silicified. The latter contain irregular siliceous nodules presumably formed by circulating hot waters and vapors associated with hot-spring and solfataric action. Under the microscope the fine beds are seen to consist of glass, crystal fragments, chiefly feldspar, quartz, biotite, and augite, and much fine unidentified material. These minerals and the associated fragments of gray obsidian indicate a composition more acidic than that of the flows. Some crystal tuffs are andesitic, however, and locally so firmly welded as to be indistinguishable from altered lava. Well-bedded pyroclastics are clearly water laid, but massive layers may have been deposited directly from the air. Scattered beds of conglomerate and tuffaceous sandstone, 2 to 6 feet thick, are associated with the water-laid ash and tuff. The conglomerate contains well-rounded stones of lava, 2 to 12 inches in diameter, in a sparse sandy matrix.

Distinctive white, gray, and red lapilli tuff in groups of beds up to 200 feet thick are more abundant than the finer layers. The white beds consist largely of pumiceous fragments an inch or less in diameter set in a tuffaceous matrix of glass and crystal shards. The red and gray layers contain varicolored scoriaceous to pumiceous fragments with a few pieces of dark dense lava and obsidian, usually less than an inch in diameter, all embedded in red or gray cinders.

Crudely stratified agglomerate and breccia are abundant in beds 50 to 75 feet thick. They consist of angular to subangular pieces of vesicular basalt and andesite up to 5 feet in diameter, embedded in a scanty tuffaceous matrix.

Andesite flows well up in the section contain angular fragments of granite and schist a few inches in diameter which have been carried up several thousand feet from the older basement rocks. The schist appears to be essentially unmodified, but the granite shows evidences of leaching and perhaps differential fusion, for it is notably cellular, frayed at the edges, and digested in appearance.

FIGURE 1. TERTIARY VOLCANICS NORTH OF WOLF CREEK
Aerial photograph looking north from over Wolf Creek. By W. A. Wood.

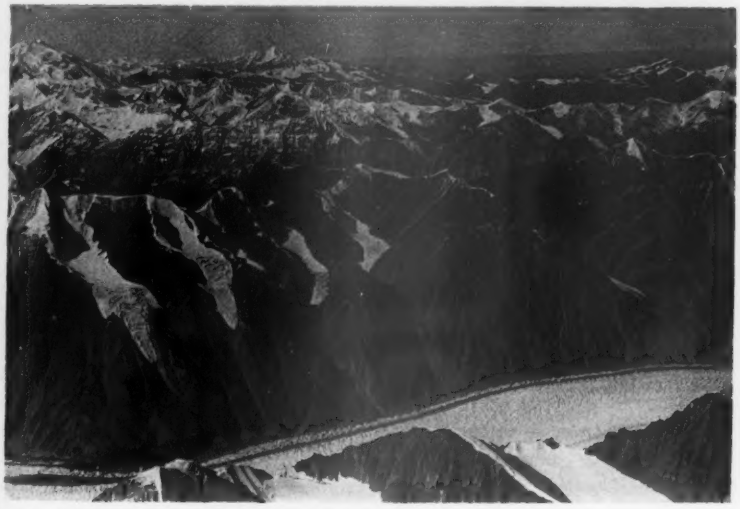


FIGURE 2. REGION SOUTH OF WOLF CREEK
Foreground rocks possibly Devonian-Mississippian metamorphics. Mt. Wood at upper left. Looking north. (Aerial photograph by W. A. Wood.)

REGION ADJOINING WOLF CREEK

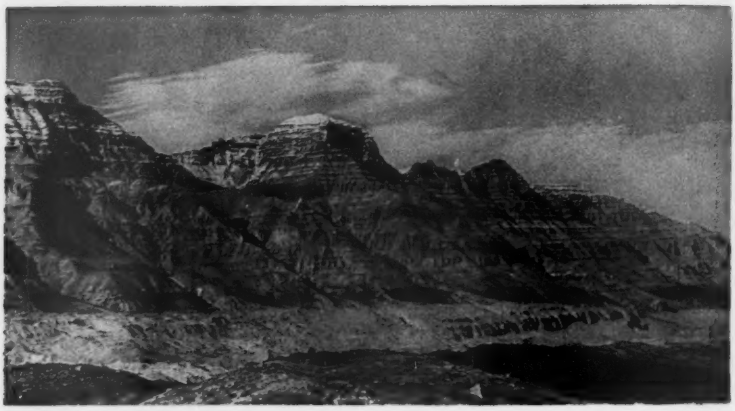


FIGURE 1. TERTIARY VOLCANICS, NORTH WALL OF WOLF CREEK
Normal section described in text is exposed on flank of central snow-capped peak. About 4600 feet of volcanics shown.

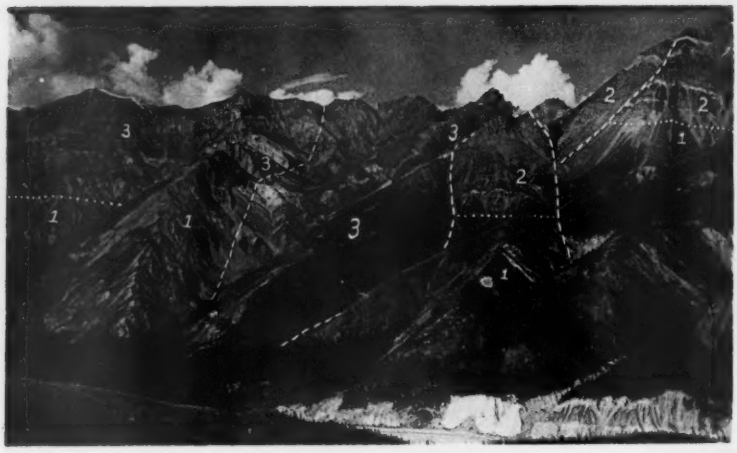


FIGURE 2. FAULTS IN NORTH WALL OF LOWER WOLF CREEK
1—Devonian-Mississippian metamorphics, 2—Permian strata, 3—Tertiary volcanics. Dashed lines are faults, dotted lines are unconformities.

VOLCANICS AND STRUCTURE

Thickness, arrangement, and structure.—The thickness of these volcanics was originally greater than the 4600 feet now exposed, for the uppermost layers have been eroded. McConnell (1906, p. 23) estimates their minimum thickness in this vicinity at 5000 feet, and 6000 to 8000 feet of similar rocks is reported in the Copper River region (Mendenhall, 1905, p. 55).

The succession of flows and pyroclastics shown in Figure 1 represents the normal Wolf Creek sequence exposed in the lower central part of the canyon (Pl. 5, fig. 1). It consists of the following major units:

Tertiary volcanics (descending order)

11. Interbedded lava and agglomerate with minor white tuff; layers at top are cliff forming	1000(?)
10. Breccia, agglomerate, lava, tuff, and ash in various arrangements, white and red lapilli tuff layers abundant and separated by lava flows 20 to 40 feet thick, a few thin conglomerate beds included	1500
9. White lapilli tuff layers constituting the most prominent unit in the sequence	200
8. Massive andesite flows	100
7. Agglomerate, lava, tuff, ash, and a few conglomerate and sandstone beds, fragmental material predominant over flows	600 to 800
6. Group of 15 to 20 thin flows each characterized by a thin vesicular zone at the base, a central layer of nonvesicular but commonly porphyritic lava, and an upper zone of highly vesicular brecciated reddish lava representing the broken and oxidized top of the flow; no evidence of extended weathering between flows except in one or two instances where it might be due to modern processes accelerated by ground-water seepage. From a distance this group of flows looks like well-bedded sedimentary rocks	200
5. White tuff bed	30
4. Cliff-forming, vesicular to dense basalt and andesite flows including layers of agglomerate and the distinctive amygdaloidal porphyritic basalt flow containing large amber phenocrysts of labradorite; many flows only 5 to 10 feet thick and notably vesicular	300 to 400
3. White porous ash	50
2. Uniform massive sodic quartz latite flows of dense and porphyritic texture	200 to 300
1. Massive flows forming vertical cliffs	200 to 300
Approximate total	4600

The amounts of pyroclastics and lava in this section are about equal. Flows predominate near the base where basalt is most plentiful. The middle third has more pyroclastics than average with white ash and lapilli tuff particularly abundant. The upper third consists about half and half of agglomerate and flows. Extrusive episodes were interrupted frequently by explosive outbursts and by short periods of quiescence during which conglomerate, sandstone, and other water-laid beds containing materials of local origin were deposited.

Origin.—These rocks are considered to be of local origin because: (1) The obsidian can hardly have come from a distant source, (2) large angular fragments in agglomerate and breccia beds indicate a vent of

moderate proximity, (3) the flows could have been supplied from local dikes and plugs of proper composition, (4) the area between peaks 9209 and 9197 is thought to be a filled volcanic crater, and (5) the thin flows on Wolf Creek could hardly have come from a point so distant as the vents of the Wrangell Range, 130 miles west-northwest, which supplied the Wrangell lavas so widely distributed in Alaska (Mendenhall, 1905, p. 54-62; Moffit, 1941, p. 133-141). Furthermore, similar volcanics in neighboring areas may also be of local origin (Cairnes, 1915a, p. 100; Capps, 1916, p. 58).

Deviations from the normal succession are found near the bend in Wolf Creek, where the lowermost beds are obsidian and white platy ash. Near point 5564 south of the canyon the basal materials are well-bedded tuffaceous silt, sandstone, and conglomerate. Agglomerate and breccia increase in abundance and coarseness westward. The well-bedded normal sequence gives way to a jumbled mass of breccia, agglomerate, welded crystal tuff, lava, and dikes in an area bounded on the east by peak 9197 and on the west by peak 9209. This area is considered to be a filled volcanic crater because: (1) The well-developed layering seen farther east disappears here, (2) coarse agglomerate beds dip 25° inward, (3) abundant coarse pyroclastics indicate excessive explosive action, (4) more noticeable alteration of the rocks suggests circulation of hot vapors and solutions, (5) lava flows thicken up to the edge of this jumbled area and end abruptly, (6) such flows also dip slightly outward as though they had formed on the lower slopes of a broad cone.

The volcanics are largely undeformed except along sharp linear flexures and faults. This is also generally characteristic in Alaska (Moffit and Knopf, 1910, p. 33; Moffit, 1930, p. 158).

The Wolf Creek flows came in part from vents and possibly also from fissures. Many of the coarse agglomerate beds may have been deposited as mudflows (Howel Williams, 1941, p. 380). No progressive change in chemical composition of materials erupted is recognized. In the eastern part of Wolf Creek mafic flows are abundant toward the base, but in the west the lowermost layers are obsidian and acidic tuff. This suggests eruptions from different vents or fissures supplied at about the same time with magmatic materials of different composition. Fully half the volcanism on Wolf Creek was explosive, and some of the most violent explosions were from vents supplied with acidic materials.

Pre-volcanic surface.—These volcanics rest on a relatively smooth erosion surface (Pl. 5, fig. 2) with a maximum observed relief of less than 200 feet. This does not compare with the 3000-foot relief, reported

by Mendenhall (1905, p. 57), on the surface beneath the Wrangell lavas. The Wolf Creek surface is more like that in the Talkeetna Mountains (Brooks, 1911, p. 104) and the Chisana-White River area (Capps, 1916, p. 34). Possibly these surfaces are not the same, for Mendenhall's appears to be Eocene, but that on Wolf Creek and elsewhere may be younger.

The pre-volcanic erosion surface on Wolf Creek is mantled by several feet of structureless extensively weathered detritus, and the rocks below are oxidized and kaolinized to depths of 30 feet. This weathered mantle consists of small angular fragments of the underlying rocks with a preponderance of quartz and less abundant schist and phyllite. The sandy and micaceous matrix contains kaolin and is cemented by brown hydrous iron oxides. This material is interpreted as a regolith developed on the erosion surface by subaerial weathering and subsequently buried by the volcanic materials. True soil is lacking except possibly in one place where several feet of reddish structureless clay overlies the coarser regolith.

Age.—The volcanics of Wolf Creek rest unconformably on older rocks including the late Mesozoic or early Tertiary intrusives. The thickness, grain size, and proximity of the late Eocene lignite-bearing beds east of Donjek River (McConnell, 1906, p. 21; Cairnes, 1915b, p. 15, 32-33) strongly suggest that they formerly extended into the Wolf Creek area. Lava and tuff interbedded in these strata may represent a preliminary phase of the volcanism, but absence of late Eocene deposits from Wolf Creek suggests erosion prior to eruption of the Wolf Creek volcanics. Thus, the base of these volcanics need be no older than Oligocene. Local deformation and dissection to a depth of 5000 feet or more prior to the later Wisconsin glaciation suggests that the uppermost layers are not younger than middle or late Quaternary.

The Wrangell lavas of Alaska (Mendenhall, 1905, p. 57) were extruded during a period extending from Eocene to Recent time. The Eocene age of the basal Wrangell lavas is confirmed by Moffit's (1930, p. 160; 1938a, p. 97-98) evidence, and Recent activity of the Wrangell vents is a matter of historical record. Capps (1916, p. 61) believes that local phases of the volcanism began in Eocene time, but the major activity was post-Eocene, a conclusion favored by Schrader and Spencer (1901, p. 52). Capps (1916, p. 60-61, 64) has further established the Pleistocene age of most of the lavas north of White River by finding them interbedded with and overlying pre-Wisconsin Pleistocene tillite. Canadian Yukon rocks similar to those of Wolf Creek are the Carmacks

volcanics (Bostock, 1936, p. 40-43) and the Newer volcanics (Cairnes, 1916, p. 42; Cockfield, 1921, p. 31-33), which are Eocene to Miocene. Presumably the Wolf Creek volcanics are at least partly equivalent to these formations and represent a fraction of the total span of Wrangell activity, but none of these names is adopted because the Wolf Creek rocks are of local origin. The Wolf Creek volcanism probably centered around the mid-Tertiary, although it could have been initiated as early as Oligocene and may have continued into the Quaternary.

STRUCTURE

Structures along Wolf Creek are relatively simple except in the closely folded Paleozoic metamorphics and in a faulted area at Wolf Creek mouth. Recognizable folds in the metamorphics are small, the largest being 100 to 200 feet from limb to limb, but these may be superimposed on larger structures. Many folds are isoclinal, and the almost universal parallelism between bedding and foliation suggests that this is the predominant habit. Most fold axes trend east-northeast, but the direction and degree of plunge is not consistent. The foliation dips 50 to 75 degrees northwest except near Wolf Creek mouth (Pl. 1). Deformation of the metamorphics was well advanced or completed prior to intrusion of the gabbro and may be tentatively dated as Pennsylvanian.

High-angle faults with apparent reverse displacements (Pl. 5, fig. 2) cut the Permian beds and the unconformably overlying volcanics in lower Wolf Creek (Pl. 1). They are divided into a minor set trending northeast and a major set trending northwest. The fault planes dip 80 to 90 degrees, and displacements are relatively small (500 to 800 feet) on all except the two easternmost faults which have throws of at least 2000 feet. Direct examination of fault planes was not possible, so the exact direction of relative movements cannot be given. An oblique, dip-slip or strike-slip movement could have produced the observed stratigraphic displacements. If these faults are essentially contemporaneous and related to the same stress, their high angle and orientation suggest an essentially horizontal shear couple acting in a northwest and southeast direction with the northeast side moving southeast. The superior development of the faults trending northwest could be attributed to their approximate parallelism with the shear couple (Fig. 2).

Since these faults cut the Tertiary volcanics, they are probably middle or late Cenozoic. It has been suggested (Russell, 1891, p. 130-131; 1892, p. 174; Spencer, 1903, p. 125, 129) that the St. Elias Range is a horst bounded on the northeast and southwest by faults showing considerable late Cenozoic displacement. Differential uplift along such

fractures would set up stresses within the St. Elias block which may well have caused the Wolf Creek faults.

The Tertiary volcanics are deformed by a sharp monocline trending east-northeast from north of Mt. Wood to lower Wolf Creek where

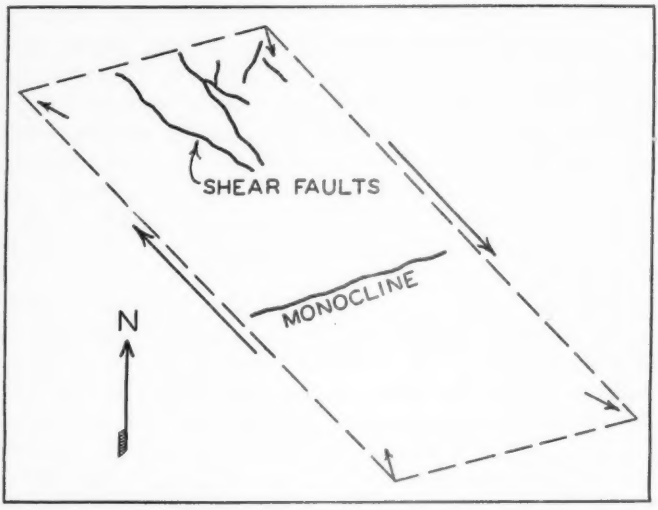


FIGURE 2.—*Analysis of middle or late Cenozoic stress-strain relations on Wolf Creek*
Southeast-northwest shear couple is postulated.

it is lost in the fault complex. The straight course of Wolf Creek from the bend eastward (Pl. 2, fig. 1) and of its north fork westward has been determined by this structure. The monocline has an upward shift on the south of several thousand feet, which accounts for the almost complete lack of volcanic rocks south of Wolf Creek. Dips up to 60° N. have been measured along this flexure. Parallelism between the monocline and foliation in the metamorphics (Pl. 1) indicates some degree of control by the latter. However, the foliation shows no evidence of warping or tilting within the monoclinal zone. This suggests a fault in the metamorphics over which the volcanics were draped in monoclinal fashion. Such a fault would be difficult to locate because it would parallel the bedding and schistosity. Location of the monocline along the margin of the massive granitic batholith is probably not accidental. Since the volcanics are displaced by this structure, it is middle or late Cenozoic and may be related to the same stresses which caused

the faults at Wolf Creek mouth, particularly if it is due to compression (Baker, 1935, p. 1502-1503; Blackstone, 1940, p. 614-616). Figure 2 is a generalized analysis of the possible stress relations which might have produced both the faults and the monocline.

SUMMARY OF GEOLOGICAL HISTORY

The earliest event recorded is deposition of at least 1000 feet of relatively pure limy sediment in a Devonian sea shallow enough to be inhabited by corals. Subsequent changes introduced much mud and some sand without interrupting deposition of calcareous material. The resulting sediments were impure limy and dolomitic beds interlayered with thin mud strata. Contemporaneous igneous activity supplied interbeds of basic volcanic debris. Deposition of several thousand feet of these mixed sediments continued through part of the Devonian or Mississippian, or both.

Diastrophism tightly compressed the Devonian-Mississippian (?) sediments into folds trending roughly east-northeast, imparted a low-grade metamorphism to the rocks, and uplifted the land. Following culmination of the deformation, but possibly contemporaneous with its closing phases, a stock of gabbro was intruded near the mouth of Wolf Creek. By some time in the Permian erosion had exposed the gabbro and developed a land surface of low relief across the older rocks. These relations suggest that the Pennsylvanian was largely a period of diastrophism, intrusion, and erosion, and this seems to have been true over much of Yukon and Alaska.

A Permian sea eventually submerged the land, and on its slowly subsiding floor many thousand feet of sand, lime, and volcanic materials were deposited. Brachiopods and bryozoans flourished in parts of the Permian sea. These strata were deformed by moderate tilting prior to intrusion of granite and quartz monzonite in upper Wolf Creek. These large intrusives are related either to the Jurassic-Cretaceous Coast Range batholith or to Cretaceous-Eocene Laramide bodies. They metamorphosed the Paleozoic rocks and produced considerable silicification and sulphide mineralization, including introduction of molybdenite.

Diastrophism associated with or subsequent to the intrusion uplifted the land and initiated erosion which exposed the granitic batholith, and developed a surface of low relief across all the older formations. A regolith was formed by weathering on this surface prior to eruption of the Tertiary volcanics which buried the area to a depth of nearly 5000 feet in basaltic and andesitic flows and more acid pyroclastics. In other regions the period between granitic intrusion and volcanism

involves deposition of Cretaceous or Eocene formations and several episodes of deformation and erosion. Evidence of these events is lacking along Wolf Creek, although Eocene coal-bearing beds crop out just to the east. During the volcanic activity extrusions of lava and explosive outbursts were about equally abundant with flows dominant in the early stages and explosions more abundant in the middle. Volcanism did not continue into the latest Pleistocene here as in other Alaskan and Yukon areas.

The volcanics were deformed by high-angle reverse faults and a monoclinal flexure in middle or late Cenozoic time, perhaps contemporaneously with phases in the last major uplift of the St. Elias block. The erosion initiated by uplift cut canyons many thousand feet deep which served as passageways for the great Pleistocene valley glaciers. The complex history of erosion, deposition, advance, and retreat of the glaciers occupying the Wolf Creek drainage is reserved for treatment in other publications. It involves several Wisconsin or partly post-Wisconsin episodes and a modern advance 100 to 200 years ago followed by retreat and stagnation interrupted by minor advances, one of which is currently in progress.

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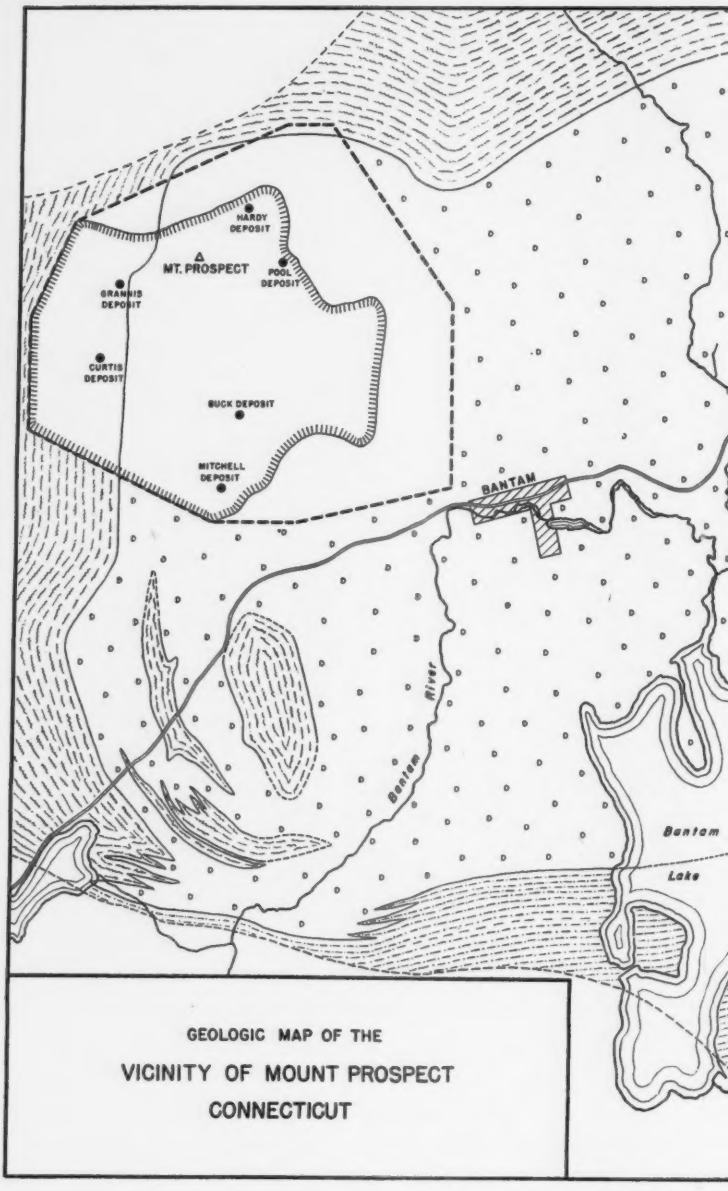

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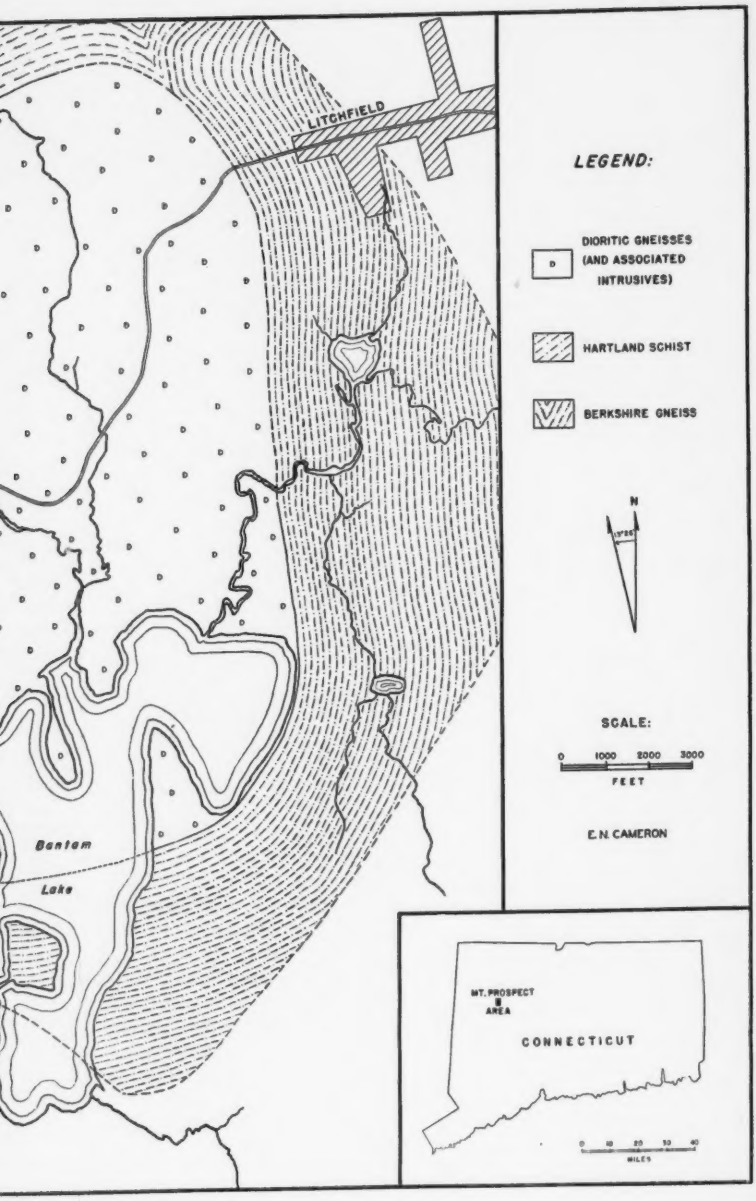

ITY OF MOUNT PROSPECT, CONNECTICUT

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ORIGIN OF SULPHIDES IN THE NICKEL DEPOSITS OF MOUNT PROSPECT, CONNECTICUT

BY EUGENE N. CAMERON

CONTENTS

	Page
Abstract	652
Introduction	652
Acknowledgments	653
Broader geologic setting of the deposits	653
Previous investigations	654
Basis and scope of the present paper	654
Geologic relations in the area of the nickel deposits	655
Preliminary statement	655
Older rocks	655
Berkshire formation	655
Diorite gneisses and associated intrusives	655
Younger basic intrusives	656
General description	656
Olivine norite	656
Quartz norite	656
Hypersthene pyroxenite	657
Dike rocks	657
Granitic rocks	657
Nickel deposits	658
Preliminary statement	658
Mitchell deposit	658
Buck deposit	658
Curtis deposit	659
Grannis deposit	659
Pool deposit	660
Hardy deposit	660
Other deposits	660
Summary of field relations	661
Description of the ores	662
Preliminary statement	662
Nonmetallic minerals	662
General description	662
Orthotectic minerals	662
Hornblende and biotite	663
Origin of hornblende and biotite	665
Hydrothermal alteration products	665
Metallic minerals	666
General description	666
Hypogene minerals	666
Supergene minerals	672

Page

Relations of sulphides to silicates 672
Relations of sulphides to hydrothermal alteration products 674
Time of sulphide deposition in the ore-bearing rocks 674
Other sulphide-bearing rocks 674
Sulphides in the gneisses 674
Sulphides in dike rocks 675
Sulphides in biotite hornblendite, biotite pyroxenite, and trachytoid norite 676
Time of sulphide deposition—temperature conditions 676
Causes of concentration of the sulphides 677
Origin and classification of the deposits 677
Comparison with other "magmatic sulphide" deposits 678
Summary of conclusions 678
References cited 679
Explanation of Plates 3–7 681

ILLUSTRATIONS

Figure Page

1. Diagrammatic summary of the textural relations of pentlandite 667
2. Pentlandite replacing pyrrhotite 668
3. Pyrrhotite partly replaced by marcasite 670

Plate Facing page

1. Geologic map of the vicinity of Mount Prospect, Connecticut 651
2. Geologic map of part of the Mount Prospect intrusive complex, Connecti ut 654
3. Sulphide-bearing rocks 682
4. Olivine norite 683
5. Photomicrographs 684
6. Photomicrographs 685
7. Photomicrographs 686

ABSTRACT

The nickel deposits of Mount Prospect are small concentrations of pyrite, pyrrhotite, pentlandite, and chalcopyrite. The ores for the most part are sulphide-rich facies of olivine norite, quartz norite, and hypersthene pyroxenite intruded in the order named, but at certain deposits wall rocks and xenoliths also contain sulphides. Field and laboratory studies indicate that the sulphides formed by fracture filling and metasomatic replacement of silicates after intrusion of a series of dikes which cut the ore-bearing rocks. The deposits are therefore regarded as hydrothermal, rather than magmatic or late magmatic as previously classified. Hornblende and biotite, developed in the host rocks prior to sulphide deposition, probably are due in considerable part to alteration during an initial stage of mineralization.

INTRODUCTION

The nickel deposits of Mount Prospect in northwestern Connecticut have been known for more than 100 years. They are small and economically unimportant but exhibit ores of a type long interesting to geologists—basic intrusive rocks containing pyrite, pyrrhotite, pentlandite, and chalcopyrite, and virtually free from characteristic hydro-

thermal alteration products. The ores therefore belong to the widespread group recognized by Vogt (1894) as the "magmatic sulphide" type, and previous investigators have classified them either as magmatic or late magmatic.

Re-examination of the nickel deposits in connection with a study of the Mount Prospect intrusive complex has revealed data bearing upon origin. The present paper records these data and discusses their significance.

ACKNOWLEDGMENTS

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To Dr. William M. Agar the writer wishes to express his lively appreciation for support and encouragement and for continued interest in progress of the work. To his wife, Mrs. Adrienne M. Cameron, the writer is deeply indebted for capable assistance in the field and in the laboratory and for most helpful criticism. To Mr. Eric L. Evans special thanks are due for care and interest taken in preparing polished specimens of a troublesome set of ores. Mr. Joseph Kubish and Mr. John Helding, of Bantam, Connecticut, extended many courtesies in the field. Professors S. J. Shand and C. H. Behre, Jr., of Columbia University, have made many helpful suggestions. Prof. Shand also criticized the manuscript.

BROADER GEOLOGIC SETTING OF THE DEPOSITS

The broader geologic setting of the deposits is shown in Plate 1. The oldest rocks are gneisses and schists of the Berkshire and Hartland formations (Agar, 1927). South and east of Mount Prospect the two formations have been invaded by gneissic diorites and associated biotite pyroxenite, biotite hornblendite, and granodiorite porphyry. Within the area marked by the broken line, these older rocks are intruded by a younger series, predominantly basic. Granitic rocks which appear to be still younger complete the assemblage of bedrock units.

The nickel deposits are associated with the younger basic series. The locations of the six principal deposits are indicated in Plate 1. Workings consist of pits and shafts now largely inaccessible. Little is known of their history except that they represent mining ventures carried on intermittently since about 1835.

PREVIOUS INVESTIGATIONS

The occurrence of sulphides on Mount Prospect was noted by Shepard (1837, p. 39) and by Percival (1842, p. 72). Later Hobbs (1906) described some of the rocks, certain sulphide-bearing rocks among them. In 1915 Howe published results of a detailed microscopic investigation of ores from the Mitchell, Buck, and Grannis deposits. He showed that these ores are sulphide-bearing rocks ranging from gabbro and norite to peridotite. He was uncertain whether the rocks are several distinct intrusives or phases of a single type. Howe concluded that the sulphides, together with associated hornblende and biotite, crystallized directly from magma and hence are magmatic in the strict sense. He attributed (p. 347) sulphide concentration to "differentiation of gabbroitic magma prior to its extrusion."

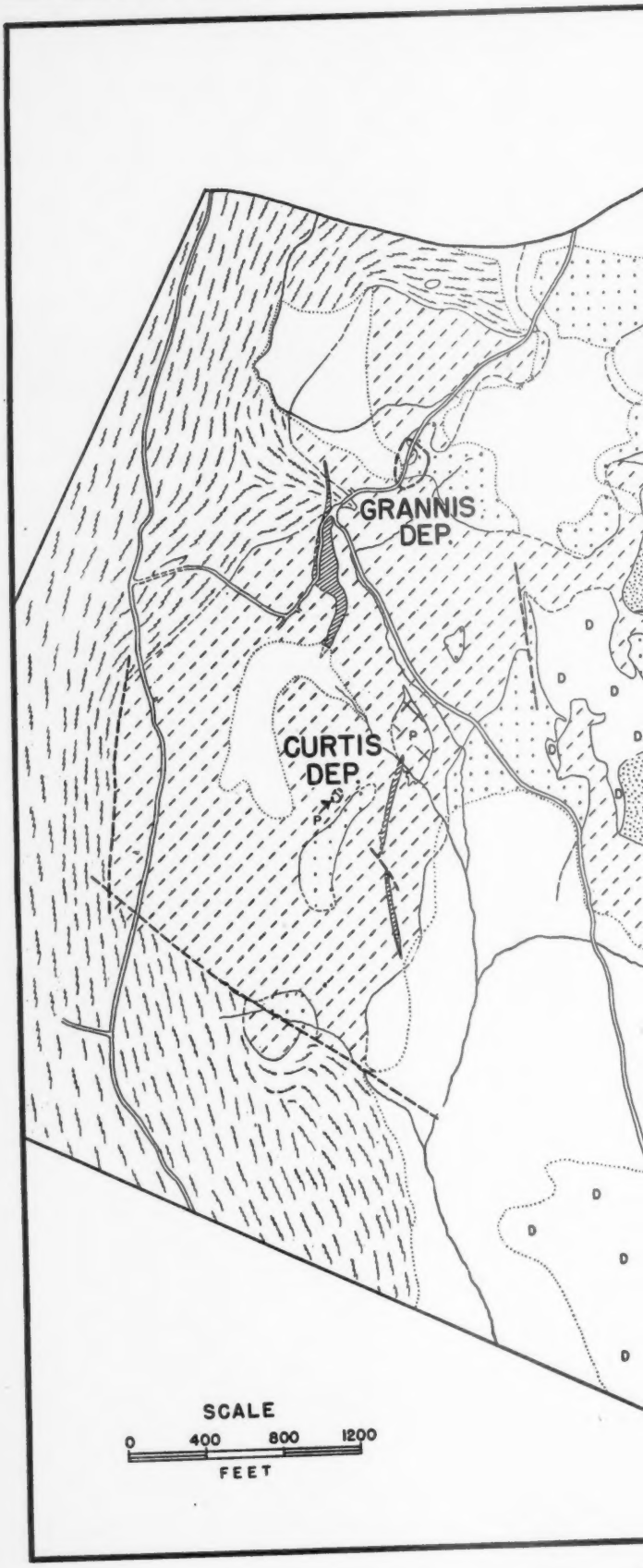
Tolman and Rogers (1916, p. 43), after examining material furnished by Howe, expressed general agreement with his findings but classified the ore minerals as late magmatic—that is, introduced since rock solidification and formed by replacement of silicates. They also reported late-magmatic hornblende and biotite and slight "post-magmatic" serpentinization of olivine. Despite their conclusions, however, Howe's work has since been cited by Lindgren (1933, p. 799) as a convincing exposition of the thesis that the sulphides are of primary magmatic origin.

Only scanty information on field relations was available to the early investigators, hence their conclusions were based largely on microscopic studies. Agar's report (1927) on the region surrounding Mount Prospect contributed much valuable information on the older metamorphics and associated intrusives and the granitic rocks, but the area of the nickel deposits was only briefly mentioned.

BASIS AND SCOPE OF THF PRESENT PAPER

The present paper is based on about 6 months of field work during 1939, 1940, and 1941. About 4 months were spent in study of the younger basic intrusives. An area of about 2¼ square miles (bounded by the hachured line in Pl. 1) was plane-tabled on a scale of 400 feet to 1 inch. The results, somewhat generalized, are shown in Plate 2. Field work has been supplemented by microscopic study of about 250 thin sections of rocks of the area and 42 polished specimens of sulphide-bearing rocks.

The petrology and structure of the Mount Prospect intrusive complex are to be discussed in a later report. In the present paper attention is restricted to the sulphide problem.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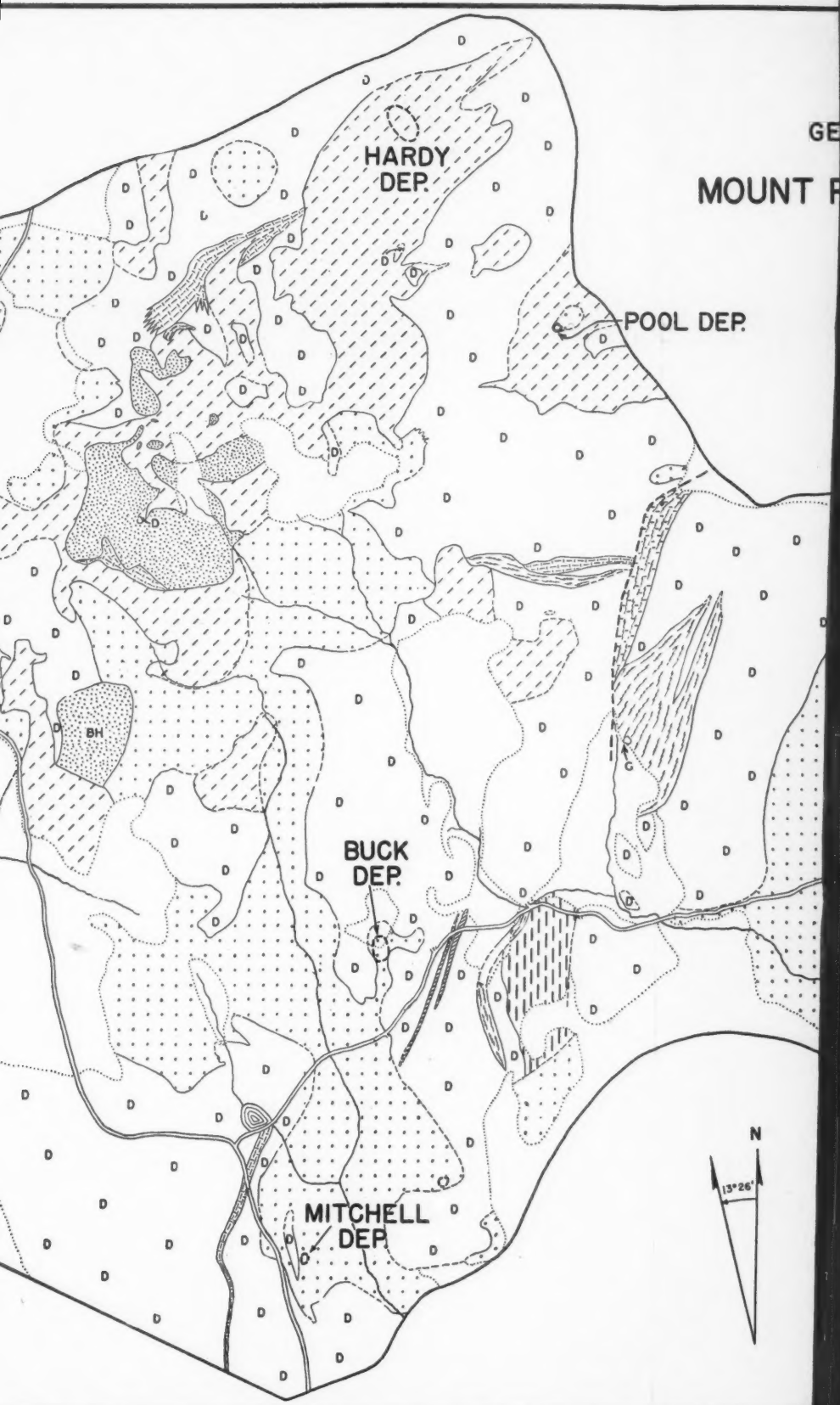
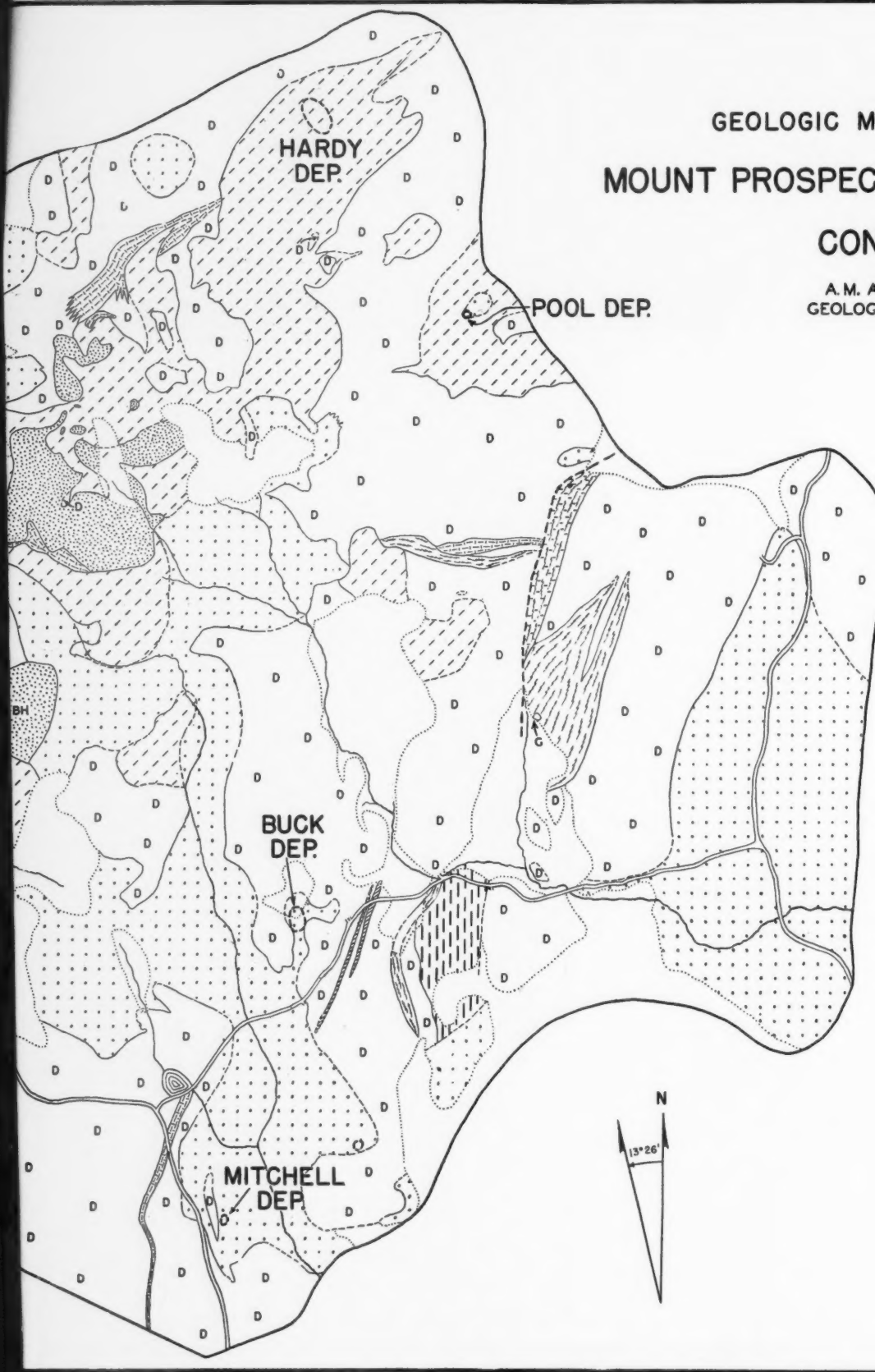


GEOLOGIC MAP OF PART OF THE MOUNT PROSPECT INTRUSIVE COMPLEX, CONNEC



OGIC MAP OF PART OF THE MOUNT PROSPECT INTRUSIVE COMPLEX, CONNECTICUT

GIC MAP OF PART OF THE
SPECT INTRUSIVE COMPLEX
CONNECTICUT

A. M. AND E. N. CAMERON
GEOLOGY BY E. N. CAMERON

LEGEND:

- COVERED
- G — GRANITE
- GRANITIZED SCHIST

YOUNGER BASIC ROCKS

- DIKE ROCKS
- HYPERSTHENE PYROXENITE
- QUARTZ NORITE
- OLIVINE NORITE
- NORITE – AGE UNCERTAIN

OLDER ROCKS

- GRANODIORITE PORPHYRY
- BH — BIOTITE HORNBLENDITE
- BIOTITE PYROXENITE
- D — DIORITIC GNEISSES
- BERKSHIRE GNEISS

DEP. DEPOSIT

---- FAULT

GEOLOGIC RELATIONS IN THE AREA OF THE NICKEL DEPOSITS

PRELIMINARY STATEMENT

The area of Plate 2 presents an extraordinary variety of rock types. For the present purpose three groups may be recognized: (1) older rocks, the Berkshire formation together with diorite gneisses and associated intrusives; (2) younger basic intrusives; (3) granitic rocks.

OLDER ROCKS

Berkshire formation.—Mica gneiss of the Berkshire formation (Agar, 1927, p. 26-28) borders the complex on the west and north and occurs as xenoliths within many of the intrusives.

The formation is variable. The prevailing type is a fine- to medium-grained contorted mica gneiss, dominantly quartz and biotite with varying amounts of muscovite, oligoclase, and orthoclase. Accessories are garnet, apatite, and sillimanite. Composition ranges from biotite-quartz gneiss to quartz-biotite schist. Lenses and layers of coarse quartz are common.

Diorite gneisses and associated intrusives.—Large portions of the area (Pl. 2) are underlain by variable diorite gneisses.

Texture ranges from fine- to coarse-grained, and composition from quartz diorite gneiss to pyroxene diorite gneiss. Foliation is usually distinct, and schistose biotite-rich and hornblende-rich facies are common. A large-scale layering accompanies the foliation, and the total assemblage is best described as a pile of alternating layers of different textural and mineralogical types. The predominant minerals, in approximate order of abundance, are andesine, hornblende, biotite, quartz, clinopyroxene, and orthoclase, but mineral proportions vary widely from layer to layer. Accessories are magnetite, pyrite, sphene, apatite, epidote, allanite, and zircon.

Much of the variation is due to invasion by massive to gneissoid granodiorite porphyry (or quartz monzonite porphyry). This rock forms numerous dikes and sills and in addition has extensively injected the diorite gneisses lit-par-lit or irregularly. Mixed rocks are abundant. Only a few porphyry masses were mapped separately from diorite gneisses.

The rock shows large microcline crystals in a medium- to coarse-grained groundmass of oligoclase-andesine, quartz, orthoclase, and biotite, with accessory pyrite, magnetite, apatite, epidote, allanite, sphene, and zircon.

On the ridge east of the Grannis deposit medium-grained massive biotite pyroxenite occurs in patches—evidently fragments of a single mass which has been brecciated and invaded. The biotite hornblendite farther south on the same ridge may be a mass originally biotite pyroxenite but now hornblendized. Probably these two intrusives are related to irregular biotite hornblendite dikes (Agar, 1927, p. 30) which cut diorite gneisses in the Shepaug aqueduct tunnel south and southeast of the

area (Pl. 2). The biotite pyroxenite is younger than the diorite gneisses but older than granodiorite porphyry, which cuts it.

The biotite pyroxenite is chiefly biotite and diopsidic clinopyroxene, with subordinate brown hornblende, plagioclase (An_{55-90}), and hypersthene (En_{74}), and accessory serpentinized olivine, apatite, magnetite, and ilmenite. Traces of pyrrhotite, pentlandite, and chalcopyrite are present.

The biotite hornblendite is medium- to coarse-grained and consists chiefly of biotite and hornblende with subordinate diopsidic clinopyroxene, small amounts of plagioclase (An_{50}), and accessory apatite, sphene, anatase (?), zircon, iron oxides, and pyrrhotite. Noteworthy are the poikilitic hornblende and biotite, the former measuring 2 to 15 centimeters in diameter in certain facies.

YOUNGER BASIC INTRUSIVES

General description.—The younger basic intrusives range from granodiorite to peridotite, but norites and pyroxenites predominate. As recognized by Hobbs (1906, p. 26) some are strikingly similar to rocks of the Cortlandt series in New York. Four members of the series are concerned in the present problem. In order of intrusion these are olivine norite, quartz norite, hypersthene pyroxenite, and dike rocks. The first three are ore-bearing. The dikes are important because they afford a means of dating sulphide deposition. The order in which the intrusives were emplaced is clearly shown by simple crosscutting relations.

Olivine norite.—The olivine norite forms three large and several small independent masses with highly irregular contacts. In part the irregularities are due to emplacement in shattered areas, with development of contact breccias. In part they are due to subsequent brecciation of the olivine norite itself, which forms angular inclusions in later intrusives. Thus, for example, it seems evident that the olivine norite patches in quartz norite south of the Curtis deposit are xenoliths derived from the large olivine norite mass to the east (Pl. 3). Outcrops in the vicinity show many smaller xenoliths, not mapped.

The olivine norite is essentially plagioclase (An_{55-70}) and laminated hypersthene (En_{75-79}) of the Bushveld type (Hess and Phillips, 1938), with varying amounts of clinopyroxene, olivine (Fo_{71-79}), hornblende, and biotite. Accessories are magnetite, ilmenite, apatite, and zircon. Spinel is present in some specimens. Most contain pyrrhotite, pentlandite, and chalcopyrite in scattered small grains, but sulphides are abundant only at the deposits. Mineral proportions vary greatly, even within a few feet, so that the rock ranges from mesotype olivine norite to peridotite. Pyroxene-rich olivine norite predominates. Texture is equally variable, from intersertal to strikingly poikilitic, and from medium-grained to extremely coarse, with feldspar and hypersthene up to 2.5 centimeters. Some phases exhibit poikilitic hornblendes .5 to 10 centimeters long. Other variations reflect assimilation or shearing but are not concerned with the present problem.

Quartz norite.—The quartz norite bodies also have outlines partly determined by brecciation of older rocks. Where well exposed, therefore,

the boundaries are locally highly irregular. Xenoliths are abundant. The large mass extending southwest from the Grannis deposit appears, from mapping of its foliation, to be floored like a steep-sided elliptical basin. Other masses appear irregular.

The quartz norite is typically medium-grained, massive to indistinctly foliated, and consists of labradorite (An_{50-62}), laminated hypersthene (En_{66-73}), and clinopyroxene, with varying amounts of hornblende, biotite, and accessory apatite, magnetite, ilmenite, and zircon. Small amounts of quartz are present in nearly all specimens. In some the texture is patchy, ranging from medium to coarse. Poikilitic hornblende and biotite mark striking variant facies. Reaction with inclusions or wall rocks is responsible for further variations, such as spinel-bearing or garnet-bearing phases. Pyrrhotite, pentlandite, and chalcopyrite, though present in most specimens, are abundant only at the deposits.

In the southeastern part of Plate 2 is a mass of partly uralitized trachytoid norite similar in many respects to quartz norite though strikingly different in texture. Field evidence suggests it is older, however, than olivine norite, and for this reason it is shown in Plate 2 as "norite of uncertain age". The sequence of mineral development in the rock is essentially similar to that shown by quartz norite.

Hypersthene pyroxenite.—In the western part of Plate 2, narrow dikes and two small irregular masses of hypersthene pyroxenite cut quartz norite.

The pyroxenite is medium-grained and composed essentially of hypersthene (En_{72-78}) and clinopyroxene. It contains small amounts of serpentinized olivine, plagioclase (An_{50}), hornblende, and biotite, and accessory magnetite, ilmenite, apatite, and sulphides.

Dike rocks.—Many dikes cut the norites and pyroxenite and the older group of rocks as well. Most are an inch or less to several feet in width. A few large ones are shown in Plate 2. The dike sequence has not been fully determined, but most are younger than quartz norite, and some at least are younger than hypersthene pyroxenite. The types involved in the problem of sulphide deposition are discussed in subsequent sections. The dikes are included in the younger basic series because they are coextensive with the olivine norite, quartz norite, and hypersthene pyroxenite.

The dikes range from granodiorite through norite to hornblende pyroxenite, dioritic types prevailing. The usual difficulties have arisen in assigning names to those which lie between gabbro (or norite) and diorite, predominantly hornblende and/or biotite with a plagioclase more calcic than An_{50}. For these the name gabbro-diorite is used, and the composition of the plagioclase is given.

GRANITIC ROCKS

Granite and granitized schist form two small masses in the area (Pl. 2), and granite pegmatites cut Berkshire gneiss west of the intrusive complex.

They are probably related (Agar. 1927, p. 34, 35) to the widespread Thomaston granite of western Connecticut. The granitic rocks are younger than diorite gneisses, biotite hornblendite, and granodiorite porphyry, which they cut south of the nickel deposit area. Their age with respect to the younger basic series is uncertain. At the Hardy deposit a garnetiferous granite aplite cuts diorite gneiss previously contact-metamorphosed by quartz norite. Near the same deposit and near the Grannis deposit quartz norite is cut by quartz-microcline-oligoclase stringers. These observations furnish somewhat dubious evidence that the granitic rocks, in part at least, postdate the basic series. Regardless of age, the granitic rocks appear to have no bearing on the sulphide problem.

NICKEL DEPOSITS

PRELIMINARY STATEMENT

The long controversy over "magmatic sulphide" deposits has shown the need for information on their field relations. For this reason, and because the deposits on Mount Prospect differ from one another, relations at the six principal deposits are described. The location and approximate extent of each are shown in Plate 2. In ensuing descriptions, by "sulphides" is meant pyrite, pyrrhotite, pentlandite, and chalcopyrite.

MITCHELL DEPOSIT

The Mitchell deposit is in the southwestern part of a large irregular olivine norite mass, near its contact with a large xenolith composed of Berkshire gneiss, diorite gneiss, and granodiorite porphyry. A shaft shows olivine norite containing smaller xenoliths of contact-metamorphosed diorite gneiss and Berkshire gneiss. Sulphides form sparsely distributed blebs, grains, and veinlets in olivine norite and in some specimens of Berkshire gneiss.

BUCK DEPOSIT

The Buck deposit is in an irregular offshoot of the Mitchell olivine norite mass. The olivine norite is variable in texture and mineral proportions, in places peridotitic or rich in coarse poikilitic hornblende. Both olivine norite and surrounding diorite gneisses are cut by norite, gabbro-diorite, and diorite dikes. Visible workings are a partly caved pit and a flooded timbered shaft. The pit exposes olivine norite cut by two hornblende gabbro-diorite dikes (plagioclase An_{60-64}), one containing sulphides.

The mineralized area is poorly exposed but evidently small. The ore is mostly olivine norite with sulphides irregularly distributed in small streaks, patches, grains, and thin veinlets. Some veinlets cut both

olivine norite and later thin norite and hornblende-biotite diorite dikes (Pl. 4, fig. 3).

One dump shows a rock composed essentially of corroded plagioclase crystals (An_{85-95}) enclosed in hypersthene, clinopyroxene, biotite, and hornblende. Abundant sulphides are present. The origin of this rock is uncertain. It seems best interpreted as derived from recrystallized and mineralized xenolith of diorite gneiss encountered undergound. No comparable rocks outcrop in the vicinity and the diorite gneiss within 15 feet of the pit described above shows no sulphides. However, since the sulphides clearly are very erratically distributed in the deposit this latter observation may not be significant.

Southwest of the Buck deposit are four shallow pits and a small quarry. The olivine norite in these excavations shows only small amounts of sulphides.

CURTIS DEPOSIT

The Curtis deposit is exposed in a shallow pit in a small hypersthene pyroxenite mass. The predominant rock is massive pyroxenite containing abundant sulphides rather uniformly distributed in small grains and aggregates. One part of the pit shows pyroxenite with small plagioclase-rock inclusions. Both inclusions and pyroxenite contain irregularly distributed sulphides, but less than the normal pyroxenite. The source of the inclusions is unknown.

Since the hypersthene pyroxenite mass is not exposed elsewhere, the extent of the sulphide-bearing phase is unknown. Other hypersthene pyroxenite masses in the vicinity show only traces of sulphides.

GRANNIS DEPOSIT

The Grannis deposit lies along the brecciated margin of a mass of olivine norite, as its contact with quartz norite. Both norites are cut by small dikes, chiefly fine-grained gabbro-diorites, diorites, and amphibolites. The result is considerable complexity, and the geologic relations indicated in Plate 2 are generalized. The quartz-norite appears to have a basin-shaped floor. There is nothing to indicate, however, that sulphide concentration is related to the floor.

Workings include a shallow pit, two flooded inclines, and an adit. Sulphides occur in olivine norite, in quartz norite, and in amphibolite, quartz-mica diorite, leucocratic quartz diorite, and garnet-bearing diorite dikes, and in all except quartz-mica diorite are found partly in thin veinlets, partly in irregularly distributed patches and grains. In quartz-mica diorite only disseminated grains are present. In olivine norite irregular masses up to 2 inches in diameter are found. Sulphides are abundant in olivine norite, which evidently was the chief ore, but are unevenly distributed. Much olivine norite is virtually barren.

The garnet-bearing diorite dikes are of special interest. Such dikes are a distinctive type widespread in the western part of the area. Near the Curtis deposit they cut both quartz norite and hypersthene pyroxenite. At the Grannis sulphide veinlets cut not only this type of dike but also later thin leucocratic quartz diorite dikes.

POOL DEPOSIT

Relations at this deposit are obscure but evidently complex. Surrounding it are diorite gneisses intruded by granodiorite porphyry and a rock which appears to be gneissoid quartz norite. The ore is chiefly variable olivine norite, which is exposed at the mouth of a flooded opening, but the dumps show sulphides in norite, diorite gneisses, and granodiorite porphyry, which must have been encountered underground. A shear zone of decayed material separates olivine norite from hornblendic norite in the roof of the opening.

Judging from the dumps the ore resembled Buck and Grannis ores in manner of sulphide occurrence.

HARDY DEPOSIT

The Hardy deposit was worked by two shafts now flooded. Neighboring outcrops show variable medium- to coarse-grained quartz norite. The dumps show this rock together with contact-metamorphosed and mineralized diorite gneiss and Berkshire gneiss xenoliths. In both xenoliths and quartz norite the sulphides are unevenly distributed in grains, aggregates, thin veinlets, and irregular ellipsoidal masses up to 5 inches in diameter. The last (Pl. 3, fig. 3) are of considerable interest. They contain abundant inclusions of silicates and may show either sharp or "fading" contacts against enclosing rocks. The masses are connected by veinlets along shear surfaces or irregular fractures.

Specimens from the dumps show quartz norite cut by a thin fine-grained gabbro-diorite dike (An_{53}). The quartz norite contains small sulphide patches connected by irregular thin veinlets, some extending across the dike. Other specimens show contact-metamorphosed diorite gneiss cut by a garnet-bearing granite aplite. The relation of sulphides to aplite is not shown.

OTHER DEPOSITS

There are shallow pits in sulphide-bearing olivine norite at two other points in the Mount Prospect area. One is 880 feet east-northeast of the Mitchell deposit (Pl. 2), along the probable margin of the Mitchell olivine norite. The other is in olivine norite 2080 feet S. 33° E. of the Buck deposit. At both pits exposures are poor. Features shown are better displayed at other deposits.

SUMMARY OF FIELD RELATIONS

Six aspects of the field relations appear significant:

(1) The deposits are closely associated with and coextensive with the younger basic series. No deposits are known outside the broken line of Figure 1. Whether epigenetic or syngenetic, the deposits seem related to the basic intrusives.

(2) The ore-bearing rocks are of several genetically distinct types: olivine norite, quartz norite, and hypersthene pyroxenite.

(3) Except possibly at the obscure Curtis and Pool deposits, the ores are local facies of intrusives which elsewhere contain only traces of sulphides. It seems unlikely that differentiation of gabbroitic magma prior to its intrusion (Howe, 1915, p. 347) could produce such concentrations.

(4) No systematic arrangement of the ores with respect to the intrusives is apparent. The Buck deposit occupies a marginal position. The Mitchell and Hardy deposits adjoin xenoliths. The Grannis deposit appears to lie along part of the floor of a quartz norite mass, but most of the ore is in the olivine norite floor, not in the quartz norite. Simple gravitative settling therefore cannot account for concentration of the sulphides. The Pool and Curtis ore bodies are so poorly exposed that their positions cannot be ascertained.

(5) Sulphides are not restricted to olivine norite, quartz norite, and hypersthene pyroxenite. Traces of sulphides are present in biotite pyroxenite and biotite hornblendite (pyrrhotite only), trachytoid norite southeast of the Buck deposit, granodiorite porphyry, diorite gneisses, and Berkshire gneiss. Sulphides occur in the last three rocks chiefly at the Mitchell, Pool, Grannis, and Hardy deposits, where they form xenoliths in, or wall rocks of, ore-bearing intrusives.

Dikes intruded into quartz norite and olivine norite at the Buck, Grannis, and Hardy deposits likewise contain sulphides. Included at the Grannis is one type which is younger than hypersthene pyroxenite. In part the sulphides occur as veinlets which cut the dikes.

(6) Except possibly at the Curtis deposit, the sulphides are irregularly distributed.

It is evident that the sulphide problem is more complex than previous investigators realized, for the presence of sulphides in a number of different rocks must be explained. From field relations it is clear that something more than orthomagmatic crystallization is involved. Orthomagmatic crystallization explains neither the sulphides in rocks older than the younger basic series nor the postdike veinlets, which require consideration especially. The veinlets raise a question; namely—To what extent is sulphide deposition due to orthomagmatic processes, and

to what extent to introduction after the host rocks solidified? An answer must be sought by study of the sulphide-bearing rocks, beginning with the ores.

DESCRIPTION OF THE ORES

PRELIMINARY STATEMENT

The olivine norite, quartz norite, and hypersthene pyroxenite, though intruded at different times, are closely similar in composition and in mineral relationships pertinent to the sulphide problem. In the following sections, therefore, the ores are discussed jointly, with comments on features peculiar to individual ores and ore-bearing rocks.

NONMETALLIC MINERALS

General description.—It is believed that three genetic groups of nonmetallic minerals can be recognized in the ore-bearing rocks. The first group includes olivine, hypersthene, clinopyroxene, plagioclase, quartz (in part), and the minor accessories apatite, spinel (in part), and zircon. These appear to be orthotectic. The second group consists of hornblende, biotite, and spinel. Where age is indicated, hornblende and biotite appear to be postorthotectic. The third and last-formed group comprises fibrous tremolite-actinolite, talc (?), sericite, chlorite, carbonate, serpentine, and quartz. These minerals are considered due to hydrothermal alteration and introduction.

Orthotectic minerals.—Olivine, hypersthene, clinopyroxene, plagioclase, and quartz (in part) show textures which appear readily explained by orthotectic crystallization involving reaction of olivine and magma to form hypersthene, and of hypersthene and magma to form clinopyroxene. Nothing in the textures requires origin by alteration, introduction, or replacement after rock solidification. For lack of evidence to the contrary, apatite and zircon are likewise classified as orthotectic.

Olivine is found only in olivine norite and hypersthene pyroxenite. It forms small rounded or corroded inclusions in pyroxenes, and occasionally in hornblende.

Hypersthene forms anhedral to euhedral crystals or granular aggregates. In olivine norite and hypersthene pyroxenite it is euhedral against plagioclase (Pl. 4, fig. 1). In some phases of quartz norite it forms euhedra surrounded by plagioclase and sometimes poikilitically enclosed; in others it fills interstices of a plagioclase lath plexus. Gradations into ophitic texture are found.

Clinopyroxene, like hypersthene, forms granular aggregates or detached crystals. Against olivine and hypersthene it is irregular and interpenetrating and often includes "corroded" grains of these minerals. With respect to plagioclase it is generally similar to hypersthene, but in a few olivine norite specimens it occurs with plagioclase in granular aggregates interstitial to coarser hypersthene-clinopyroxene aggregates.

Plagioclase likewise varies in habit. In olivine norite and hypersthene pyroxenite it is interstitial with respect to pyroxenes with the exception noted above. In

olivine norite it sometimes forms large irregular crystals poikilitically enclosing pyroxenes. In quartz norite it ranges from anhedral, where interstitial or poikilitic, to subhedral or euhedral where enclosed in pyroxenes.

QUARTZ interstitial to pyroxenes and plagioclase is found in small amounts in nearly all specimens of quartz norite. Though obviously late, it is assigned to the orthotectic stage for lack of contrary evidence. Traces of quartz are also present occasionally in the other two ore-bearing rocks, but always intergrown and apparently contemporaneous with hydrothermal minerals.

APATITE and rarer ZIRCON form inclusions in other silicates. In quartz norite acicular apatite is particularly abundant in plagioclase. SPINEL (pleonaste) is present in norite contaminated with Berkshire gneiss.

Hornblende and biotite.—In most specimens of the ore-bearing intrusives, hornblende and biotite accompany the orthotectic minerals. As Howe pointed out, the origin of hornblende and its relations to the sulphides are important, since hornblende is known to have formed in some ore deposits by hydrothermal or pneumatolytic alteration. The same is true of biotite.

Biotite and hornblende textures in the ores are complex. They indicate that hornblende and biotite, in the order named, formed largely or wholly by metasomatic replacement of orthotectic minerals in rocks already solid. Findings of the writer therefore agree more closely with conclusions of Tolman and Rogers (1916, p. 43) than with those of Howe (1915).

HORNBLENDE occurs in the following ways:

(1) As irregular poikilitic crystals 2 millimeters to 3 centimeters long including few or many "corroded" or "rounded" pyroxene and plagioclase crystals, or molded around hypersthene. This hornblende closely resembles the "poikilitic hornblende" described by Williams (1886, p. 31-34) from Cortlandt Series rocks and recently discussed by Shand (1942, p. 416).

(2) As irregular grains, usually interstitial.

(3) As pseudomorphs after hypersthene and clinopyroxene, developing at their contacts with plagioclase. In such cases, the pyroxene outlines usually are preserved only in the earliest stages of hornblende development.

(4) In intricate parallel intergrowth with clinopyroxene.

(5) As irregular-walled veinlets along fractures and microfaults in hypersthene, clinopyroxene, and plagioclase.

(6) As irregular partitions separating the units of feldspar and pyroxene crush-mosaics.

The first two textures might be due either to orthotectic crystallization involving magmatic resorption or reaction, or to replacement of pre-existing minerals in a rock already solid. However, the latter explanation seems more consistent with the fact that every stage in the development of hornblende by replacement of close-packed pyroxene or pyroxene-plagioclase aggregates seems to be shown.

If hornblende preceded plagioclase, hornblende pseudomorphs after pyroxene might also be explained by magmatic reaction. However, hornblende clearly postdates plagioclase wherever age relations are indicated. The pseudomorphs appear due to replacement which frequently involved reaction between rhombic pyroxene and clinopyroxene, or between pyroxene and plagioclase, in the presence of water

(*cf.* Shand, 1942, p. 417, 418). The last three textures can only be due to replacement guided by weaknesses caused by slight crushing of rocks already solid. Inasmuch as every gradation exists among the six textures described, it is difficult to escape the conclusion that all the hornblende developed by the same mechanism.

If hornblende developed by postsolidification replacement, several other features find ready explanation: frequent size disparity between hornblende and earlier-formed minerals, erratic distribution of hornblende, and striking antipathetic variation of hornblende and plagioclase. For example, in one half of a thin section of olivine norite from the Buck deposit (Pl. 5, fig. 4) euhedra and subhedra of hypersthene and clinopyroxene are poikilitically enclosed in plagioclase. In the other half, the matrix material is coarse hornblende, poikilitic like the plagioclase, but with pyroxene inclusions which are markedly irregular. The hornblende also contains "corroded" inclusions of plagioclase and along the boundary between the two areas is in irregular interpenetrating contact with plagioclase, in places veining it. It seems apparent in such cases that hornblende developed by replacement of plagioclase and pyroxene.

To summarize, much of the hornblende shows textural relations which indicate metasomatic replacement of rocks already solid. Other textures are compatible with orthotexis but seem better explained by replacement.

Howe (1915, p. 333) reported "instances where the development of hornblende rims" on pyroxene "has been followed by later growth of pyroxene and this in turn followed by more hornblende." Although pyroxene-hornblende intergrowths are common the writer has found none which support Howe's statement in any of 62 thin sections of rocks from the three deposits Howe studied.

The hornblende ranges from brown to pale green. The full color range may be shown in a single thin section or even patchily distributed within a single large crystal. Usually there is no indication that color variations correspond to differences in time or origin (*cf.* Shand, 1942, p. 415), but there are certain exceptions. Some hornblende in olivine norite shows irregular brown cores grading into pale-green margins. The reverse is not found. This suggests that the green hornblende formed in part after the brown hornblende. Also, in a few specimens from the Buck deposit fractured brown hornblende has been healed by pale-green hornblende, granular or in optical continuity. In other specimens, however, the color difference is clearly related to the minerals from which hornblende has formed: brown hornblende after clinopyroxene, spinel-bearing green hornblende at plagioclase-hypersthene contacts.

BIOTITE, brown to brownish-green, occurs in the following ways:

(1) As irregular flakes.

(2) As ragged crystals poikilitic toward plagioclase, pyroxene, or hornblende (Pl. 6, fig. 2).

(3) As crystallographically oriented flakes in pyroxenes, plagioclase, and hornblende.

(4) In pyroxene and hornblende as groups of irregular patches in optical orientation with one another and usually with surrounding biotite.

(5) As irregular-walled veinlets traversing pyroxenes, plagioclase, and hornblende.

(6) On hornblende or pyroxene as dactylytic fringes projecting into plagioclase (Pl. 6, fig. 1). Fringes are usually undeformed even if extending into crushed feldspar.

It seems evident that in large part at least biotite developed after hornblende, by postorthotectic replacement of earlier minerals.

Origin of hornblende and biotite.—Textural features indicate that hornblende and biotite formed largely or wholly by metasomatic replacement after host-rock solidification, but do not show when. Such textures might result either from deuteric attack or hydrothermal alteration. All that is needed is infiltration of hot water carrying small amounts of potash. Field evidence indicates that postorthotectic hornblende and biotite formed in the Mount Prospect rocks at more than one time and by more than one process. Near the Grannis deposit, for example, quartz norite adjacent to certain diorite dikes is heavily hornblendized and biotitized for as much as 18 inches from the contacts. Quartz norite xenoliths in the dikes are even more strongly affected. Such development of hornblende and biotite is evidently an exomorphic effect of the dikes. The hornblende is typically dark green to green brown but is not distinct from some hornblende in ore-bearing rocks far from dikes.

It is noteworthy that in the dike contact zones hornblende and biotite replace pyroxenes and plagioclase in the same manner as in the ore-bearing rocks elsewhere, except that the poikilitic hornblendes formed are small. Such occurrences support the hypothesis that hornblende and biotite in the ores developed by postorthotectic replacement. However, it seems clear that while textures may reveal mechanism of development, they do not show the cause. The bulk of the hornblende and biotite in the ore-bearing intrusives appears unrelated to dikes, but beyond this no conclusions seem justified.

Hydrothermal alteration products.—Fibrous tremolite-actinolite, sericite, talc (?), chlorite, serpentine, carbonate, and quartz, intergrown in various combinations, are present in small amounts in all three ore-bearing intrusives. These minerals have formed at the expense of biotite and older silicates. Serpentine, containing fine-granular magnetite, has formed in olivine norite and hypersthene pyroxenite at the expense of olivine. The most abundant hydrothermal mineral is tremolite-actinolite, which occurs largely as pseudomorphs after hypersthene or clino-pyroxene.

The relation between hornblende and tremolite-actinolite has been examined with particular care. Usually there is a sharp contrast between the two types of amphibole in a given thin section. However, in olivine norite from the Buck deposit a few crudely zoned amphibole crystals have been found. The crystals have irregular brown hornblende centers which fade outward into green hornblende and then pass into fibrous colorless amphibole penetrating adjacent plagioclase. Similar gradations

are found in quartz norite. Such features suggest that hornblende development was part of a continuous alteration process which terminated locally in development of fibrous amphibole and typical hydrothermal minerals.

METALLIC MINERALS

General description.—Both hypogene and supergene metallic minerals are present in the ore-bearing rocks. The hypogene minerals are magnetite, ilmenite, hematite, pyrite, pyrrhotite, pentlandite, and chalcopyrite. Marcasite and an iron-nickel mineral (violarite or bravoite?) not yet identified are also present, but are considered supergene.

Hypogene minerals.—The hypogene minerals fall into two groups—oxides and sulphides.

MAGNETITE AND ILMENITE occur as minute inclusions in pyroxenes, hornblende, and serpentine (magnetite only), and as grains or aggregates .1 to 2 millimeters in diameter scattered sparsely among silicates or included in sulphides. For the most part the inclusions in silicates appear related to the crystallization of the hosts or to subsequent changes in them (magmatic reaction, exsolution, development of hornblende and biotite, serpentinization of olivine). The granular magnetite and ilmenite appear to have formed after biotite, in places occurring as undeformed veinlets or tongues along cleavages of deformed biotite or hornblende. They appear older than the sulphides, which include "corroded" magnetite and ilmenite grains and occasionally vein them. Magnetite grains are euhedral to anhedral: ilmenite is anhedral. Ilmenite is the predominant oxide. Many grains contain minute HEMATITE inclusions oriented parallel to (0001) of ilmenite. Some grains show lamellar twinning apparently pyramidal.

Although granular magnetite and ilmenite are commonly present in the ores, a close genetic relation between oxides and sulphides is not apparent. Oxides occur in phases of the ore-bearing intrusives which contain no sulphides. Furthermore, the oxides are lacking in sulphide veinlets, except where the veinlets cut or pass adjacent to oxides.

Granular PYRRHOTITE in anhedra .05 to 25 millimeters in diameter is the predominant sulphide.

The larger crystals occasionally show slender wedge-shaped twin lamellae, apparently pyramidal. Etch and microchemical tests of the pyrrhotite reveal no unusual characteristics. Most appears homogeneous, but some from the Hardy deposit shows two components distinguishable in poorly polished specimens due to a slight difference in hardness. The two components are in parallel intergrowth, the harder appearing as patches, partial rosettes, or oriented minute lenses and blebs in the softer. The structure suggests the α-pyrrhotite–β-pyrrhotite intergrowths which Ramdohr and Schneiderhohn (1931, p. 136) and Scholtz (1936, p. 154) have suggested

are due to exsolution. Comparison with pyrrhotite from Waterfall Gorge, Insizwa, shows, however, that the contrast in color, pleochroism, and anisotropism commonly displayed by Insizwa pyrrhotite components is lacking in Hardy ore. Only the hardness difference makes it possible to recognize that two components exist, and many grains, perhaps due to orientation, appear homogeneous. Conclusions

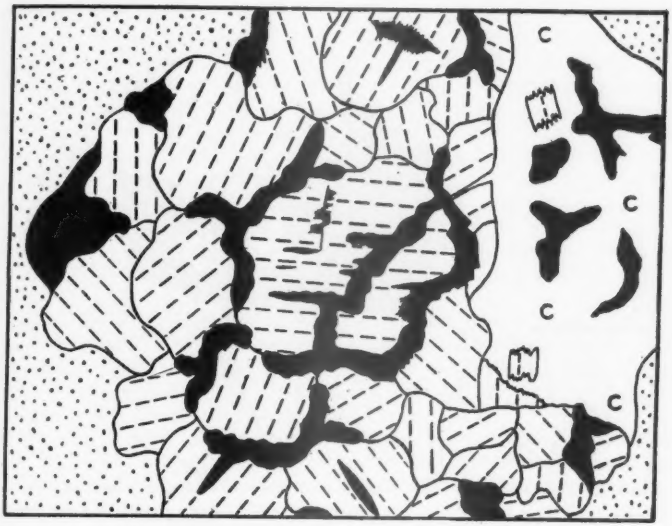


FIGURE 1.—*Diagrammatic summary of the textural relations of pentlandite*
Pentlandite, black; pyrrhotite, dashed line symbol, direction of lines indicating trace of (0001); chalcopyrite, C; silicates, stippled.

regarding the composition, distribution, and relative abundance of the two components have not been reached.

PENTLANDITE is irregularly distributed in the ores but is present in appreciable amounts in all specimens examined.

Pentlandite shows no unusual properties. Microchemical tests for cobalt were negative, indicating that it is very low or absent (Short, 1940, p. 194-197). Cleavage is usually distinct, especially where emphasized by development of supergene minerals. Pentlandite is always intergrown with or in contact with pyrrhotite, except where selective replacement of pyrrhotite has left pentlandite embedded in chalcopyrite (Pl. 3, fig. 2). The textural relations of pentlandite to pyrrhotite are summarized in Figure 1. Pentlandite occurs: (1) as single grains or aggregates at contacts of pyrrhotite with silicates; (2) as stringers along contacts of pyrrhotite grains, in the larger sulphide aggregates sometimes forming interrupted networks which outline grains of pyrrhotite; (3) as wedge-shaped or irregular-walled offshoots into pyrrhotite [parallel (0001)] from the networks (Pl. 3, fig. 1); (4) as wedges in pyrrhotite, usually parallel to (0001); (5) as irregular veinlets traversing large

pyrrhotite grains, in no case extending beyond the boundaries of a single pyrrhotite crystal; (6) as flamelike inclusions in pyrrhotite parallel to (0001). For descriptive purposes the six habits are distinguished, but it must be emphasized that there are gradations between them. The inclusions are often developed from fractures in pyrrhotite. Against pyrrhotite, the larger pentlandite grains occasionally show

FIGURE 2.—*Pentlandite replacing pyrrhotite*

Magnetite, slant-ruled; silicates, stippled. Except for area A, which seems to have been rotated slightly, the areas of pyrrhotite are in optical continuity and appear to be parts of a single original grain. Olivine-norite, Pool deposit. Tracing from photomicrograph, × 100.

partial octahedral form. Some larger pentlandite areas contain what appear to be "corroded" remnants of pyrrhotite optically parallel to neighboring pyrrhotite (Fig. 2).

Although with the one exception noted, pentlandite does not occur independently of pyrrhotite, the converse does not seem to hold true. Even allowing for the two-dimensional limitations of polished sections, many pyrrhotite grains apparently neither contain nor are in contact with pentlandite. Also, on the scale of the polished sections there seems to be no fixed ratio of pyrrhotite to pentlandite from place to place.

Pentlandite-pyrrhotite textures similar to those described have been reported from Sudbury, the Gap mine, the Alexo mine, Insizwa, and elsewhere. Opinions differ concerning their meaning. Campbell and Knight (1907, p. 365), Uglow (1911, p. 675), Tolman and Rogers (1916, p. 14, 15, 43), Wandke and Hoffman (1924), and others have interpreted the textures to indicate successive deposition, pentlandite following pyrrhotite and according to some writers replacing it. Buddington (1924) concluded that in some ores pentlandite formed in part contemporaneously with pyrrhotite, in part later. Dickson (1904, p. 61) and Howe

(1915, p. 338, 339) reached no definite conclusions. Schneiderhöhn (1922, p. 178; 1929, p. 213, 240) and Van der Veen (1924, p. 41) attributed the textures to unmixing of a pentlandite-pyrrhotite solid solution, with segregation of pentlandite to the pyrrhotite grain borders. Van der Veen pointed out that in Sudbury ores the pentlandite "veinlets" typically are not true crosscutting structures but lie along pyrrhotite intergrain contacts. Subsequently Newhouse (1927) showed experimentally that pentlandite is markedly soluble in pyrrhotite at high temperatures. From synthetic solid solutions of the two minerals he obtained intergrowths which duplicated many textures shown by natural pentlandite inclusions in pyrrhotite. Newhouse suggested that in natural ores the intergrain pentlandite "veinlets" were separated from pyrrhotite due to immiscibility during the liquid stage. Upon crystallization of pyrrhotite some adjacent pentlandite would be taken into solid solution. The pentlandite remaining liquid would lie between pyrrhotite crystals and crystallize there, or if fracturing intervened, in veinlets in magnetite and silicates. Unmixing of the solid solution would later yield pentlandite inclusions in pyrrhotite. The uneven distribution of the pentlandite could thus be explained, together with the fact that on a small scale the pentlandite-pyrrhotite ratio seems to vary in Sudbury ore, though fairly constant for the ores as a whole.

By heating Sudbury ores Hewitt (1938) dissolved pentlandite in pyrrhotite. Upon cooling, specimens heated above 800° C. showed unmixing and segregation of pentlandite to pyrrhotite grain borders. Instead of homogeneous pentlandite, however, oriented intergrowths of pyrrhotite in pentlandite, rare in natural ores and absent before heating, were produced.

Both Newhouse and Hewitt considered it improbable that all the pentlandite is due to exsolution and called upon liquid immiscibility to account for separation of all (Newhouse) or part (Hewitt) of the pentlandite aside from the inclusions. Similarly Schneiderhöhn and Ramdohr (1931, p. 123) distinguished between "independent" pentlandite and "exsolved" pentlandite inclusions. Scholtz's interpretation of Insizwa pentlandite (1936, p. 161) involves liquid immiscibility, unmixing from solid solution, and independent crystallization of pentlandite.

The experimental data furnished by Newhouse and Hewitt demonstrate that exsolution of pentlandite from pyrrhotite is possible, but the theories of unmixing presented by them, by Schneiderhöhn (1929), by Van der Veen (1924) and by Scholtz (1936, p. 161, 202-207) appear to assume that the sulphides were segregated and crystallized during an orthomagmatic or late-magmatic stage. However, what if deposition is due to introduction of sulphides after host-rock consolidation, under hydrothermal conditions? Reasons are presented below for believing that this is true for Mount Prospect ores. Under such conditions can liquid immiscibility be called upon for the segregation which both Newhouse and Hewitt require to account fully for pentlandite distribution?

In view of this uncertainty, the possibility that metasomatic replacement of pyrrhotite by pentlandite could produce the textures shown in the Mount Prospect ores must be considered. Let us suppose solutions containing nickel ion to be percolating through a pyrrhotite aggregate. If deformation produces no throughgoing fractures, the solutions must penetrate largely along intergrain contacts or along the parting of pyrrhotite. If reaction between solutions and pyrrhotite takes place, could the intergrain networks of pentlandite develop?

Mount Prospect ores seem to offer a partial answer to this question. Weathered

specimens show bladed marcasite developing from pyrrhotite (Fig. 3), in a manner described from many ores (*cf.* Schneiderhöhn and Ramdohr, 1931, p. 139). This change undoubtedly is a replacement of pyrrhotite by marcasite. Replacement proceeds in large part centripetally from intergrain contacts. In pentlandite-free areas the result in intermediate stages is a marcasite network enclosing corroded

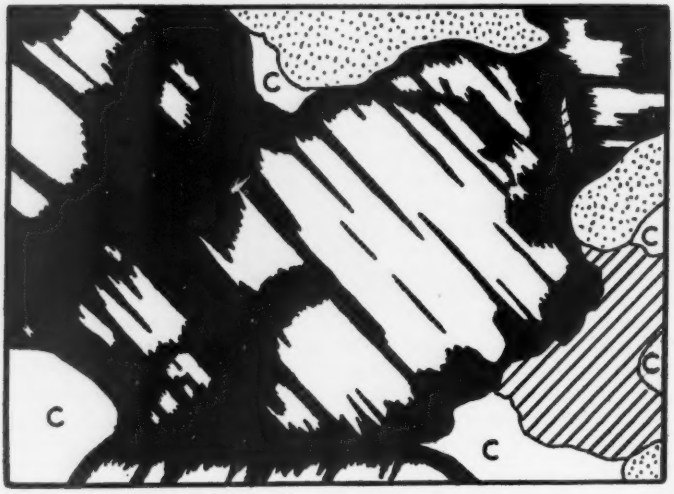


FIGURE 3.—*Pyrrhotite partly replaced by marcasite*

Pyrrhotite (blank) in an advanced stage of replacement by marcasite (black). Pentlandite, slant-ruled; chalcopyrite, C; silicates, stippled. Quartz norite, Hardy deposit. Tracing from photomicrograph, × 146.

pyrrhotite residuals. Through-going marcasite veinlets are not developed. Allowing for the difference between the granular habit of the intergrain pentlandite and the bladed habit of the marcasite, there seems to be no essential difference between this pattern and that of hypogene pentlandite-pyrrhotite. The comparison can be carried further. In part marcasite occurs as isolated blades parallel to (0001) of pyrrhotite. Like pentlandite inclusions, these are lenticular or wedge-shaped in section. In other places marcasite has developed from fractures as fringelike or brushlike patches extending into pyrrhotite along (0001). They, like the pentlandite inclusions, are irregularly distributed.

The comparison fails to hold in strongly weathered ore, for marcasite replacement can be seen in all stages from incipient development to complete pseudomorphism of pyrrhotite, whereas pentlandite replacement is not shown beyond the intermediate stages. Is this fatal to a hypothesis of metasomatic replacement by pentlandite, or is it due to the limited amount of nickel ion available?

The marcasite-pyrrhotite relation is certainly no proof that metasomatic replacement produced the pentlandite-pyrrhotite textures, but it does seem to demonstrate the possibility.

The origin of the pentlandite remains an open question. Unfortunately laboratory experiments furnish no criteria for distinguishing exsolution from metasomatic replacement effects in ores of the Mount Prospect type. The writer considers that metasomatic replacement better explains: (1) the irregular distribution of pentlandite, (2) veinlets which cut pyrrhotite grains (Fig. 1), (3) corroded pyrrhotite inclusions in pentlandite. However, only one definite limit to interpretation can be recognized. In Hardy and Pool ores pentlandite occurs entirely isolated from pyrrhotite and embedded in chalcopyrite. The chalcopyrite has plainly developed in pyrrhotite-pentlandite intergrowths by selective replacement of pyrrhotite along its basal parting. If pentlandite formed by exsolution from pyrrhotite, unmixing must have preceded chalcopyrite deposition. This is true both for the small inclusions and for the coarser pentlandite.

Study of polished sections of Mount Prospect ores revealed no evidence of the "second generation" of pentlandite reported from many similar ores. There is no more ground for believing that the inclusions developed after the coarser pentlandite than for concluding that the marcasite inclusions in pyrrhotite formed at a different time from the peripheral marcasite.

Chalcopyrite is irregularly distributed through all sulphide-bearing rocks examined, usually subordinate to pyrrhotite, approximately equal to pentlandite, but it is the predominant sulphide in a few specimens of Hardy ore.

It usually forms anhedral grains or groups at intergrain contacts of other sulphides, or of sulphides with silicates. It also occurs as independent grains, aggregates, and veinlets, and in veinlets associated with pyrrhotite and pentlandite. Its marginal relations usually indicate little of age relations, but in places it forms veinlets crossing pyrrhotite and rarely pentlandite, or has formed by selective replacement of pyrrhotite along the pyrrhotite parting. In a few specimens corroded pyrrhotite residuals are found in chalcopyrite.

There seems little doubt that chalcopyrite formed after pentlandite, largely by replacement of pyrrhotite.

Though not previously reported, pyrite is a minor constituent in all Mount Prospect ores. The deposits therefore belong to the large group (Schwartz, 1937, p. 33) in which pyrite and pyrrhotite are associated. Pyrite forms isolated crystals embedded in other sulphides and thin veinlets which cut all other hypogene sulphides. The isolated crystals, where embedded in pyrrhotite, are sometimes euhedral but usually appear "corroded." This type thus appears to have formed before pyrrhotite. The veinlet pyrite is clearly younger than chalcopyrite, though presum-

ably hypogene. Schwartz (1937, p. 35) reported a similar case of two generations of pyrite from Fierro, New Mexico.

Supergene minerals.—Marcasite and an unknown nickel sulphide are abundant in ores from all six deposits. Both are considered supergene, for in suites of specimens the degree of replacement of pyrrhotite and pentlandite by these minerals can be correlated with degree of weathering. Limonite invariably accompanies them.

MARCASITE: In partly weathered specimens pyrrhotite shows partial to complete replacement by a lamellar metallic mineral (Fig. 3). In least-weathered specimens it has the usual properties of marcasite, but with increase in degree of weathering its color changes from pale yellow (white against pyrrhotite) to gray, and its hardness, anisotropism, and reflectivity decrease. Similar variations are reported by Schneiderhöhn (1931, p. 196). The gray mineral is not distinct from typical marcasite, into which it grades imperceptibly.

UNKNOWN NICKEL SULPHIDE: A cream to violet-gray mineral replaces pentlandite partly or wholly in many specimens (Pl. 3, fig. 2). Replacement follows pentlandite cleavages or irregular fractures, giving veinlet networks with irregular pentlandite remnants filling interstices. In some specimens the former presence of pentlandite is recognizable only from the perfect pseudomorphs remaining.

Alteration of pentlandite to a secondary sulphide has been described by many investigators, who have reported the products as polydymite, violarite, or bravoite. The Mount Prospect mineral does not fit any published description exactly. Possibly more than one mineral is represented. Detailed description awaits further study.

RELATIONS OF SULPHIDES TO SILICATES

Relations of sulphides to silicates (other than hydrothermal silicates) are much the same in all three ore-bearing rocks; hence it is possible to discuss the rocks as a group. Discussion is further facilitated because the principal hypogene sulphides—pyrrhotite, pentlandite, and chalcopyrite—participate in each textural arrangement in the rocks—that is, no one shows unique relationships. By "silicates" in the ensuing paragraphs is meant olivine, pyroxenes, plagioclase, hornblende, and biotite.

The sulphides occur as disseminated minute grains and aggregates; as masses 1/16 inch to 3 inches in maximum diameter, generally in irregular contact with surrounding silicates and containing silicate inclusions; as interstitial grains or patches molded around silicates; and as veinlets. These habits are distinguished for convenience of description, but there are all gradations among them. In any specimen of ore, the sulphides are irregularly distributed.

The minute disseminated grains and aggregates commonly occur at silicate contacts or along fractures in silicates. In the few places where they seem to be true inclusions, examination of surrounding grains usually suggest their development along contacts above or below the plane of the section. The larger masses frequently contain silicate inclusions which appear corroded or rounded (Pl. 6, figs. 3, 4).

The interstitial sulphides were considered magmatic by Howe (1915, p. 339, 340), who recognized that the "corroded," "embayed" outlines of silicate inclusions suggested replacement but concluded that magmatic resorption prior to sulphide deposition could explain such features. In the writer's opinion, "interstitial" habit is not a basis for discriminating between orthotexis and other processes of development in such rocks as these. Resorption plus interstitial crystallization can produce such textures, although if the sulphides are the final product of crystallization one wonders what became of the resorbed silicate material. But the same textures can be produced in a polymineralic rock by selective replacement, particularly where intersertal and poikilitic textures prevail (Pl. 4, figs. 1, 2). The same ambiguity is inherent in irregular boundaries, "embayed" margins, and "corroded" outlines.

More significant in the writer's view is the passage of interstitial sulphides into veinlets which traverse the rocks or penetrate cleavage cracks (Pl. 4, fig. 2; Pl. 5, fig. 1; Pl. 7, figs. 1, 2). Veinlets of unfractured sulphides cut all silicates, including biotite and hornblende, and ramify through shear zones and crushed areas, especially in feldspars. The veinlets have the same composition as the "interstitial" aggregates, show the same textures and paragenetic sequence. No veinlet can be found which crosscuts an interstitial mass, although veinlets may extend in opposite directions from a mass along what obviously were once parts of a single fracture. The irregular masses and interstitial aggregates appear to be local expansions of the veinlets due to selective replacement of the walls. Replacement is often indicated by the unmatched walls of the veinlets. Portions of some thin sections are small-scale breccias, in which sulphides cement and partly replace unicrystalline or polycrystalline rock fragments.

From these observations two conclusions are drawn: (1) Only one period of sulphide deposition is represented; (2) the obvious control exerted upon deposition by fractures, microfaults, microshear zones, and mineral cleavages indicates introduction and deposition after the host rocks solidified. Fracture filling and replacement predominated, the latter erratic and often highly selective. The writer thus prefers the unequivocal evidence of veinlets and the like to the dubious interstitial and inclusion relationships. His findings with regard to the mechanism of deposition therefore agree with those of Tolman and Rogers (1916, p. 43).

The relation of the sulphides to hornblende and biotite is especially important, since these minerals are postorthotectic and may be hydrothermal. The time relationship is repeatedly shown in thin sections. Sulphides vein these two minerals or form undeformed tongues projecting

along the cleavages of hornblende and biotite crystals (Pl. 7, fig. 4) which are frequently bent or broken.

The relation of sulphides to hornblende has been examined with particular care. Hornblende has formed partly by replacement of clinopyroxene in parallel intergrowth, and it is conceivable that the fractures might be older than the hornblende which they appear to cut. In such intergrowths hornblende and pyroxene have their *c* axes and plane of symmetry in common, but the hornblende cleavage is the usual amphibole type, not an inherited pyroxene cleavage. When veinlets cross basal sections of hornblende, whether brown or green, they have patterns controlled in part by the amphibole cleavage. Sulphides in contact with the infrequent fibrous amphibole have not been found.

RELATIONS OF SULPHIDES TO HYDROTHERMAL ALTERATION PRODUCTS

Fibrous amphibole, serpentine, talc (?), chlorite, sericite, carbonate, and quartz are sporadic in their occurrence, minor in amount, and apparently unrelated to the ore minerals. The hydrothermal minerals are rarely in contact with sulphides, and their time relation to mineralization is not shown. The sole exception is that in a specimen from the Buck deposit carbonate fills minute fractures in sulphides and is evidently later. However, this carbonate may be supergene.

TIME OF SULPHIDE DEPOSITION IN THE ORE-BEARING ROCKS

In the ores all that is definitely indicated is that sulphide deposition followed postorthotectic development of hornblende and biotite. Nothing in the textures indicates whether the sulphides were derived from lower, still uncrystallized portions of the present rock masses deuterically, or from an outside source. Evidence bearing on this point has been obtained, however, from other sulphide-bearing rocks of the area and is presented in succeeding sections.

Howe (1915, p. 335) found the sulphides most frequently associated with hornblende and biotite. Since, however, hornblende and biotite are present in most specimens of olivine norite, quartz norite, and hypersthene pyroxenite whether sulphide-bearing or not, the "association" may not be significant. Wherever age relations are shown the sulphides definitely postdate both silicates, but the time interval involved is unknown.

OTHER SULPHIDE-BEARING ROCKS

SULPHIDES IN THE GNEISSES

The occurrence of sulphides at certain deposits in wall rocks and xenoliths composed of Berkshire gneiss and diorite gneiss is of special interest, because they present no problem as to whether the sulphides have been introduced. Sulphides have been found in the gneisses at numerous scattered points in the area of Plate 2, but are abundant only

adjacent to ore-bearing facies of the younger basic intrusives. Sulphide textures in these rocks therefore afford a check on the possibility that sulphide textures in the ores developed by introduction and replacement. Comparison is facilitated because some diorite gneisses, though dissimilar in texture and structure, closely resemble quartz norite in composition.

A pyroxene diorite gneiss from the Pool deposit dumps is composed of andesine (An_{40}), in part antiperthitic, green hornblende, brown biotite, subordinate hypersthene and clinopyroxene, minor quartz, apatite, magnetite, and zircon, and irregularly distributed sulphides. Sulphides occur partly as veinlets, partly as small disseminated and interstitial masses, partly as larger masses containing rounded or corroded silicate inclusions (Pl. 7, fig. 3)—relations strikingly similar to those in the ore-bearing rocks. Moreover, sulphide deposition was preceded by development of hornblende and biotite from plagioclase and pyroxenes, partly as veinlets, partly by irregular replacement of crushed areas. Although the presence of biotite and hornblende in this rock as well as in the ore-bearing intrusives might suggest a relation between the two silicates and the sulphides, the age relations cast doubt upon this.

SULPHIDES IN DIKE ROCKS

The sulphide-bearing dikes cutting olivine norite or quartz norite at the Buck, Grannis, and Hardy deposits include amphibolites, norites, gabbro-diorites, quartz-mica diorite, garnet-bearing diorite, and leucocratic quartz diorite. Wherever age relations are indicated in the dikes sulphides are later than silicates, including biotite and hornblende, and occur as disseminated grains and aggregates, sometimes along shear zones; as irregular masses; and, except in the quartz-mica diorite, as veinlets which in some cases cut both the dikes and the adjacent ore-bearing rocks (Pl. 4, fig. 3; Pl. 5, figs. 2, 3). These dikes are mostly thin (½ inch to 2 inches wide), and polished slabs have been prepared from some which show wall rock, dike, and crosscutting veinlets. The veinlets unquestionably postdate both wall rock and dike. If the veinlet sulphides were deposited later than the bulk of the sulphides in the ordinary ores, then these specimens should show evidence of two periods of sulphide deposition. Microscopic examination shows that in each slab the veinlet sulphides:

(1) Are the same as the sulphides in the adjacent wall rock,

(2) Have the same textural relations, and

(3) Occur in the same paragenetic sequence. Although fracturing obviously preceded deposition of the veinlets, nowhere in the wall rocks are "early" sulphides cut by "late" sulphides. At the Grannis deposit veinlets occur in olivine norite, quartz norite, and in three different types

of dikes cutting these rocks. Polished slabs cut from one specimen show veinlets cutting across the contacts of quartz norite with successively intruded thin garnet diorite and leucocratic quartz diorite dikes. If more than one period of sulphide deposition is represented, it seems strange that reversals of paragenetic sequence are lacking in the quartz norite.

(4) Where connecting with the usual patches of sulphides, do not crosscut them. The patches appear to be merely local expansions of the veinlets.

The postdike veinlets appear to have formed at the same time as the sulphides in the dike wall rocks. Evidence of successive mineralizations or of redistribution of sulphides after dike intrusion is totally lacking.

Hornblende and biotite are present in most dikes, in part replacing feldspar and pyroxene, but in considerable part apparently orthotectic. The relation between postorthotectic development of these minerals and sulphide deposition is uncertain.

SULPHIDES IN BIOTITE HORNBLENDITE, BIOTITE PYROXENITE, AND TRACHYTOID NORITE

Sulphides in biotite pyroxenite, biotite hornblendite, and trachytoid norite are later than any of the silicates, appear to have formed by introduction and replacement, and probably formed at the same time as those in the ores. It seems hardly likely that introduction of sulphides would be confined to the immediate vicinities of the ore deposits.

TIME OF SULPHIDE DEPOSITION—TEMPERATURE CONDITIONS

When sulphide deposition took place is not precisely known. Sulphides at the Grannis, Hardy, and Buck deposits are at least younger than five types of dikes which cut olivine norite and quartz norite. Of these, garnet-bearing diorite and leucocratic quartz diorite are younger than hypersthene pyroxenite. Deposition, therefore, followed intrusion of hypersthene pyroxenite. In view of this and because the various ores are strikingly similar, probably the sulphides are all epigenetic, although at the Curtis, Pool, and Mitchell deposits, since no dikes cut ore, direct evidence is lacking. The dikes are coextensive with, and therefore seem genetically related to, the ore-bearing rocks. However, except for the norite and certain gabbro-diorite dikes they differ markedly from the ore-bearing rocks in mineral composition and must be interpreted as later differentiates from a common subjacent magma reservoir. The sulphide-bearing dikes are members of a series which includes at least nine different types. The sequence of intrusion is not fully known, but no postsulphide dikes have been found. Evidently, therefore, deposition

took place late in the magmatic epoch represented by the younger basic intrusives, probably after intrusion of the dike sequence was complete.

The dikes are either fine- or medium-grained and except for the amphibolites and leucocratic quartz diorite show fine-grained chill borders. At the Grannis deposit, for example, a garnet diorite dike is chilled against quartz norite and also against an older chilled norite dike. It would seem from this that: (1) an appreciable interval elapsed between consolidation of the ore-bearing rocks and sulphide deposition, and (2) the temperature during deposition was appreciably below magmatic temperatures.

CAUSES OF CONCENTRATION OF THE SULPHIDES

The forms of the deposits cannot be definitely determined. The causes of concentration of the sulphides are likewise not clearly indicated. The prominence of sulphide veinlets shows that fracturing was involved, but to what extent it controlled concentration is uncertain. Related to this problem is the close association of the sulphide deposits with the younger basic intrusive masses. Satisfactory information on these points is lacking.

ORIGIN AND CLASSIFICATION OF THE DEPOSITS

The sulphide deposits are magmatic in the sense that they appear related to the younger basic series. However, they are not magmatic or even late magmatic in the sense of being syngenetic with any of the rock masses now exposed and related to their individual crystallization periods. The sulphides appear to have had a deeper source, and probably the genetic relation between them and the intrusives is that of separate differentiates from a common magma reservoir. Agar (1930) reached a similar conclusion concerning the origin of pyrrhotite, pentlandite, and chalcopyrite in amphibolitized pyroxenite at Torrington, Connecticut.

If hornblende and biotite are due to alteration early in the mineralization epoch, the deposits could well be classified as hypothermal. This interpretation seems the most reasonable, but since hornblende and biotite antedate the sulphides it cannot be proved. The status of hornblende and biotite is an important matter. It is frequently argued that the "fresh" condition of ores such as those of Mount Prospect is evidence against hydrothermal deposition. The writer questions whether any rocks in which hornblende and biotite show textures like those described here can be regarded as "fresh", unless these minerals demonstrably were not developed during the mineralization epoch. Knopf and Anderson (1930) have shown that at Engels, California, development of hornblende

and biotite preceded ore development, and the two minerals have been reported elsewhere as high-temperature alteration products.

The origin of hornblende and biotite has a significant bearing also on the origin of the granular magnetite and ilmenite. The oxides have been shown to be younger, in part at least, than hornblende and biotite, but older than the sulphides. If hornblende and biotite were developed early in the mineralization process, oxides were introduced into the ores during an intermediate stage.

COMPARISON WITH OTHER "MAGMATIC SULPHIDE" DEPOSITS

The deposits invite comparison with other pyrrhotite-pentlandite-chalcopyrite concentrations in basic intrusives and tempt one to ask whether deposits of this group classified as magmatic or late magmatic may not in some cases have formed at later stages. Although it is not justifiable to draw general conclusions from a single group of deposits, one point seems worth emphasizing. Textural relations and paragenesis in the Mount Prospect ores are closely similar to those reported from other "magmatic sulphide" deposits. There is the same postorthotectic development of hornblende and biotite and traces of unmistakable hydrothermal minerals. (*Cf.* Kerr, 1924, p. 376.) The sulphides are late in the paragenetic sequence and replace prehydrothermal silicates. Only the fortunate circumstance that the ore-bearing rocks of Mount Prospect were intruded by dikes before sulphide deposition, makes the epigenetic origin of the sulphides clear. Microscopic studies of "magmatic sulphides" afford no basis for distinguishing deuteric (late magmatic) minerals from high-temperature hydrothermal minerals. It would therefore seem that classification of these deposits as epigenetic or syngenetic can only be made on the basis of satisfactory field evidence.

SUMMARY OF CONCLUSIONS

The nickel deposits of Mount Prospect are concentrations of pyrite, pyrrhotite, pentlandite, and chalcopyrite. They occur within masses of olivine norite, quartz norite, and hypersthene pyroxenite intruded in the order named. The sulphides entered after solidification of the intrusives, by fracture filling and by metasomatic replacement of silicates. At the same time small quantities entered older rocks, especially where these rocks adjoin, or form xenoliths within, the intrusives.

Prior to sulphide deposition hornblende and biotite replaced plagioclase and pyroxenes in the intrusives and in the older rocks as well. This followed consolidation of the intrusives and in considerable part may represent hydrothermal alteration during a preliminary stage of mineral-

ization. Small quantities of undoubtedly hydrothermal minerals also formed locally, but no genetic relation between them and sulphide deposition is apparent.

The ores are cut by small dikes which range from granodiorite to pyroxenite. Sulphide deposition followed intrusion of quartz diorite, diorite, gabbro-diorite, and amphibolite dikes belonging to this series. No dikes are known which are younger than the sulphides, and probably the sulphides were not introduced until dike intrusion was complete.

Both the dikes and the sulphide deposits are coextensive with the ore-bearing intrusives and appear to represent later differentiates from the same magmatic reservoir. The sulphides therefore are genetically related to the ore-bearing intrusives but are not magmatic or late magmatic in the sense of being related to their individual crystallization periods. The deposits are epigenetic and appear best classified as high-temperature hydrothermal.

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EXPLANATION OF PLATES

PLATE 3.—SULPHIDE-BEARING ROCKS

FIGURE 1.—Stringer of unknown supergene nickel sulphide (gray, minutely fractured), pseudomorphous after pentlandite, along contact of two pyrrhotite (white) grains. Offshoot of stringer extends into one pyrrhotite grain. Black, silicates. Olivine norite, Pool deposit. Reflected light. × 91.

FIGURE 2.—Pseudomorphs of limonite and an unknown supergene nickel sulphide (light gray areas) formed by alteration of granular pentlandite embedded in chalcopyrite (white). A single remnant of pentlandite is shown slightly to the left of the center. Dark gray and black, silicates. Quartz norite, Hardy deposit. Reflected light. × 81.

FIGURE 3.—Slab showing cross section of irregular mass of sulphides (light) developed at intersection of fractures in contact-metamorphosed diorite gneiss (dark). Dark spots in the sulphide mass are silicate inclusions. Hardy deposit. Natural size.

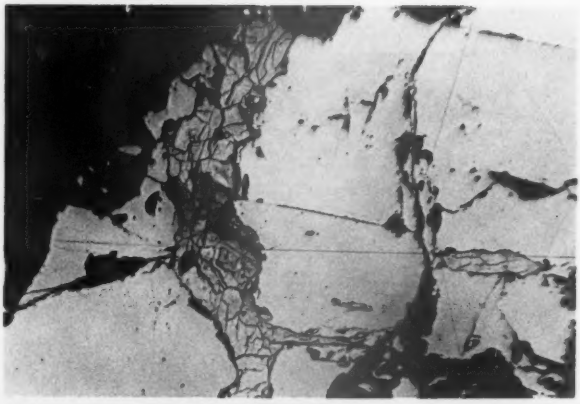
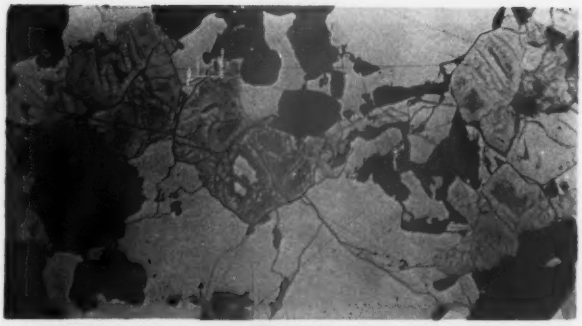


FIGURE 1



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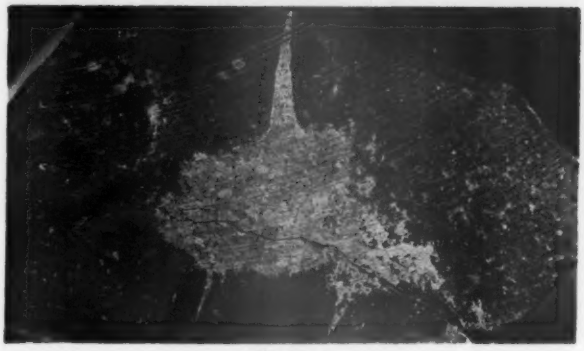


FIGURE 3

SULPHIDE-BEARING ROCKS

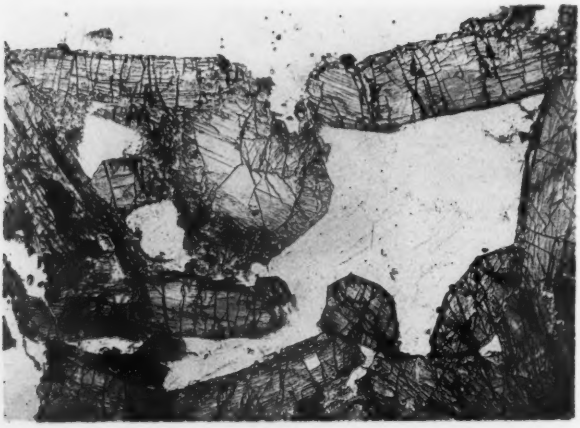
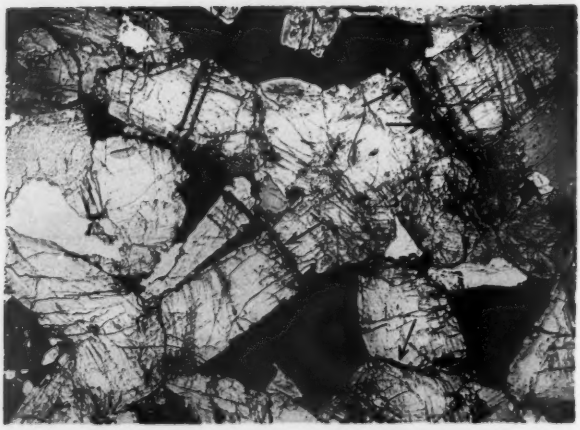


FIGURE 1



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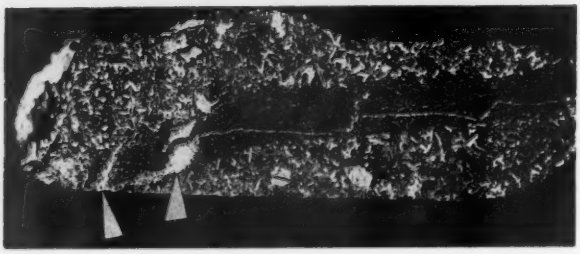


FIGURE 3

OLIVINE NORITE

PLATE 4.—OLIVINE NORITE

FIGURE 1.—Olivine norite, showing euhedral hypersthene crystals (gray, high relief) embedded in plagioclase (white). Buck deposit. × 8.

FIGURE 2.—Same specimen as Figure 1. Interstitial feldspar largely replaced by sulphide aggregates (black) which are connected by veinlets traversing intervening fractured hypersthene crystals. Three of the veinlets are indicated by arrows. × 8.

FIGURE 3.—Coarse olivine norite cut by fine-grained diorite (dark central east-west band) which is in turn cut by a veinlet of sulphides (shown by index at extreme left). Two irregular aggregates of sulphides have developed in the dike (to the right of the veinlet) partly at the expense of a presulphide plagioclase veinlet which cuts the dike. Numerous white spots in olivine norite are likewise sulphides (indicated by black arrows). Buck deposit. × 9.

PLATE 5.—PHOTOMICROGRAPHS

FIGURE 1.—Veinlet of sulphides (black) cutting peridotitic phase of olivine norite. Sulphides also form numerous minute veinlets and irregular patches in the rock adjacent to the main veinlet. Buck deposit. × 9.

FIGURE 2.—Olivine norite cut by diorite dike (central NW-SE band). Sulphides (black) extend across the dike contacts. At the right edge of the section part of a cross-cutting irregular veinlet is shown. Buck deposit. × 3.

FIGURE 3.—Irregular replacement veinlet of sulphides (black) cutting fine-grained diorite dike. Threadlike veinlets extend outward from the main veinlet into the adjacent rock. Buck deposit. × 10.

FIGURE 4.—Olivine norite. Top shows euhedral to subhedral hypersthene crystals (gray, high relief) poikilitically enclosed in plagioclase. In lower left, plagioclase has been replaced by hornblende (nearly black) intergrown with biotite (light gray). Hypersthene crystals enclosed in hornblende appear corroded. Buck deposit. × 6½.

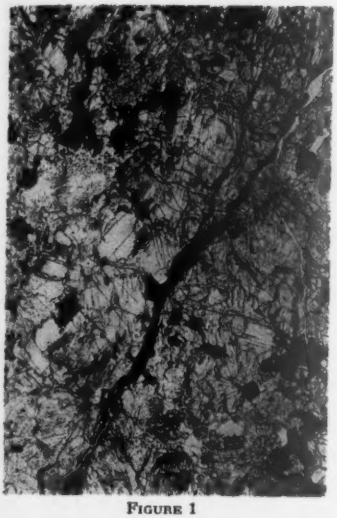
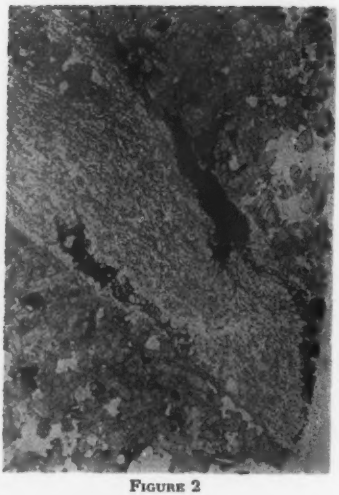


FIGURE 1



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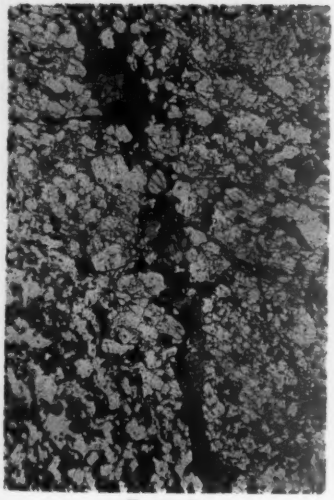

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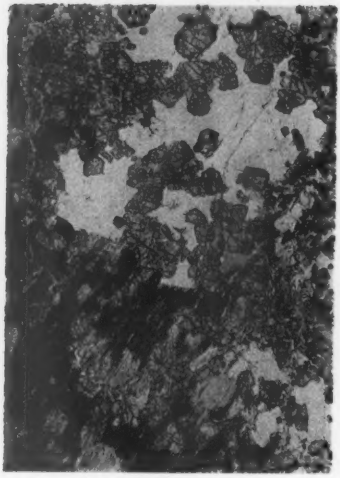

FIGURE 4

PHOTOMICROGRAPHS

FIGURE 1

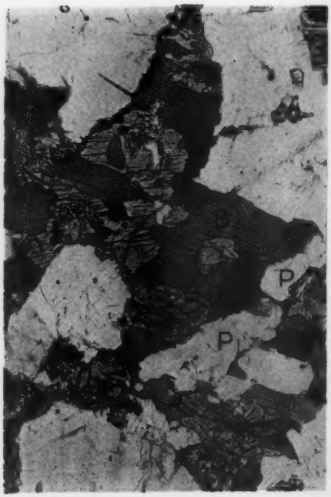

FIGURE 2

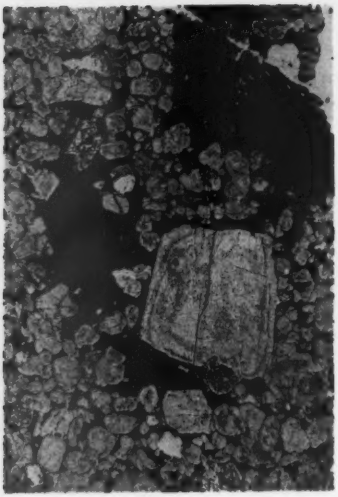

FIGURE 3

FIGURE 4

PHOTOMICROGRAPHS

PLATE 6.—PHOTOMICROGRAPHS

FIGURE 1.—Biotite (needles, gray patches with single cleavage) penetrating and replacing fractured and strained plagioclase (nearly white) and hornblende (mottled gray, high relief). Arrow points to veinlet of biotite along fractures in large plagioclase crystal. Quartz norite, Hardy deposit. × 42.

FIGURE 2.—Part of a spongy crystal of biotite (B) formed by replacement of plagioclase (white) and hornblende (gray, higher relief). P and P_1 are parts of a single plagioclase crystal microfaulted and subsequently veined and partly replaced by biotite. The hornblende contains corroded pyroxene inclusions. Quartz norite, Hardy deposit. × 33.

FIGURE 3.—"Interstitial" sulphides in the marginal part of an irregular sulphide mass. Sulphides, black—hypersthene and clinopyroxene, gray. Grannis deposit. × 6.

FIGURE 4.—"Interstitial" sulphides in the marginal part of an irregular sulphide mass. Sulphides black, plagioclase white, hypersthene and clinopyroxene gray. Grannis deposit. × 10.

PLATE 7.—PHOTOMICROGRAPHS

FIGURE 1.—Fractured feldspar (white) veined and partly replaced by sulphides (black). Gray areas are pyroxenes, hornblende, and biotite little affected except by peripheral corrosion. Quartz norite, Hardy deposit. × 7.

FIGURE 2.—Same thin section as Figure 1. Irregular mass of sulphides (black) developed largely by replacement of fractured feldspar (white). Incipient stages of replacement are shown above and to the left of the main mass. Gray areas are pyroxenes, hornblende, and biotite, little affected. × 7.

FIGURE 3.—Irregular mass of sulphides developed by replacement of pyroxene diorite gneiss. Sulphides black; plagioclase white; pyroxene, hornblende, biotite, shades of gray. Note rounding of partly replaced feldspar inclusions. Pool deposit. × 5.

FIGURE 4.—Sulphides (black) penetrating and replacing biotite (low relief, single cleavage) along cleavages. Note partly replaced flakes included in sulphide mass. Hornblende (H) and pyroxene (mottled, high relief) form scattered areas along the periphery of the figure. Olivine norite, Buck deposit. × 27.

FIGURE 1

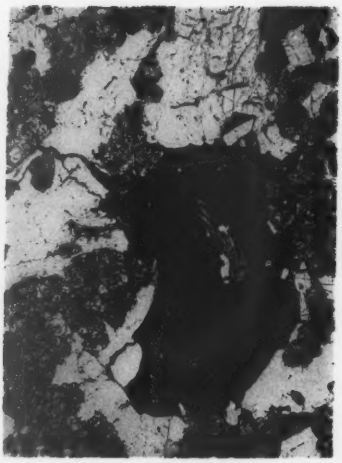

FIGURE 2

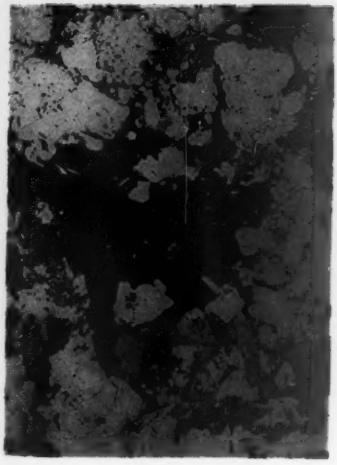

FIGURE 3

FIGURE 4

PHOTOMICROGRAPHS

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LATE PALEOZOIC FAULTING IN WESTERN NEWFOUNDLAND

BY FREDERICK BETZ, JR.

CONTENTS

	Page
Abstract	687
Introduction	688
Acknowledgments	689
Earlier observations on faulting	690
Evidence for zone of weakness	692
General statement	692
Southwestern coast	693
Middle Barachois River to Grand Lake	693
Grand Lake	695
Little Barachois River to Deer Lake	695
Deer Lake	695
Humber Valley	696
White Bay	696
Summary	698
Possibly associated faults	698
General statement	698
Grand Lake to Halls Bay	699
Little Barachois River to lower Humber River	699
Western coast north of Bonne Bay	700
Nature of faulting	700
Northward prolongation of zone of weakness	701
Age of faulting	703
Summary	704
References cited	705

ILLUSTRATIONS

Figure	Page
1. Map of Newfoundland	690
2. Geological sketch map of Newfoundland	691
3. Structure sections	694
4. Structure sections	702

ABSTRACT

A zone of structural weakness, characterized by thrust faulting, extends across Newfoundland from the southwestern coast to White Bay on the north. Studies in recent years in the Bay St. George area, at Grand Lake, and in White Bay make possible a discussion of the age and nature of the faulting. It is clearly marked in the Bay St. George area on the southwestern coast, where igneous and metamorphic rocks of the Long Range are in thrust contact with Carboniferous. In the interior, at Grand Lake, a high-angle fault separates Paleozoic lavas from Carboniferous sediments. Extensive thrust faulting in the Carboniferous paralleling the western

side of the bay, dissimilarity of the sections on the opposite sides of the bay, and topography point to the presence of the zone in the White Bay region. A pronounced lowland connects Grand Lake with White Bay; deformation of the rocks in the lowland, as well as its continuity with the White Bay trough, suggests the presence of the zone of weakness.

The faults, wherever observed, dip at relatively high angles eastward and strike north-northeast. Carboniferous rocks are involved in all the areas mentioned. The zone of weakness is believed to be late Paleozoic, probably representing a feature of Appalachian orogeny.

Faulting of similar nature has been discovered on Cape Rouge and Cape Fox peninsulas on the eastern coast of the Northern Peninsula, where Carboniferous and Ordovician are in thrust relationship. It is suggested that the remarkably straight coast of the Northern Peninsula may represent the locus of a northward prolongation of the transinsular zone of weakness.

Other possible connections of fault lines on the northwestern coast of Newfoundland and in Notre Dame Bay with the main zone are also discussed.

INTRODUCTION

In Newfoundland, at the northern extremity of the Appalachian Mountain System, thick accumulations of sediments, ranging from Cambrian into Pennsylvanian, give evidence that this was the site of geosynclinal downwarping during the Paleozoic. Recent studies, including extensive mapping by the Geological Survey of Newfoundland, have shown with increasing clarity that the orogenic episodes recognized in eastern North America also played an important part in the geological history of Newfoundland.

On structural and stratigraphic grounds, the classification of the island of Newfoundland as an integral part of the Northern Appalachians, or of the geological unit termed Greater Acadia by Schuchert (1930, p. 703), is generally accepted. The only divergent opinion of recent date is expressed by Wilson (1939, p. 281), who describes Newfoundland tentatively as a province of the Canadian Shield.[1]

On the geological map of Newfoundland compiled by Howley in 1907, the greater part of the island is shown as pre-Cambrian. A sketch map by Snelgrove (1938, Pl. II) indicates the distribution of pre-Cambrian and Paleozoic rocks in western Newfoundland and in the Avalon Peninsula to be much the same as on the older map. However, marked differences appear in the region extending eastward from the Long Range to a line running from near Fortune Bay on the south coast to Bonavista Bay (Figs. 1, 2). Most of Howley's "Laurentian & undetermined" is designated by Snelgrove as "Acidic intrusives and infolded schists, etc. Probably Paleozoic in large part." Furthermore, the large area around Notre Dame Bay mapped as "Serpentines, diorites, dolerites, etc." by

[1] A more detailed treatment of the geological features of the island as part of the Appalachian Mountain System by A. K. Snelgrove, Government Geologist of Newfoundland, will be found in a volume of *Geologie der Erde: Geology of North America*, now in press.

Howley is now known to consist of Paleozoic sediments and volcanics with granitic intrusions and a small amount of serpentine.

Still more rocks hitherto considered pre-Cambrian or of an undetermined age have been reclassified on the basis of recent work.[2] Cambrian, Ordovician, and Devonian strata have been discovered on the south coast in the Fortune Bay area, where as late as 1937 the exposed rocks were held to be almost exclusively pre-Cambrian.

As based on the newer mapping, the tendency is to give increased recognition to Paleozoics throughout Newfoundland. As a result, attention is focused on the Paleozoic history of the island, especially with regard to the periods of deformation.

Schuchert and Dunbar (1934, p. 12-13) state that four periods of orogenic activity have affected Newfoundland—Middle Ordovician, Taconic, Acadian, and post-Pennsylvanian (presumably Appalachian). Caledonian orogeny, for which suggestive evidence is presented by Heyl (1936, p. 18-19), can probably be included.

Areal reports published by the Geological Survey of Newfoundland since 1935 show that indirect evidence must be used in most cases for assuming and dating periods of deformation. Exposures of fault planes and fold axes are scarce and small. The large faults are often postulated because of topographic expression—long, narrow valleys or embayments. The time of faulting (and folding) must be determined with reference to rocks that have been tentatively dated.

It is the purpose of this discussion to consider a zone of structural weakness, characterized by thrust faulting, that extends northward across Newfoundland from the southwestern coast. The faulting is thought to be late Paleozoic, probably a feature of Appalachian orogeny. The zone to be discussed was first postulated as a fault line by Murray 75 years ago and has been mentioned in the later general literature. The inadequate appreciation of late Paleozoic deformation in Newfoundland has been due largely to limited knowledge of the Carboniferous rocks. Although parts of the zone are included in more recent areal studies, the evidence to prove it a major structural feature of Newfoundland has not been gathered previously.

ACKNOWLEDGMENTS

Particular thanks are due Dr. A. K. Snelgrove, Government Geologist of Newfoundland, who was largely responsible for the writer's interest in the subject of this paper and who examined the manuscript with great care.

[2] Unpublished information of the Geological Survey of Newfoundland.

The writer is deeply indebted to Dr. Helgi Johnson, of Rutgers University, for continued interest and kind assistance on numerous occasions during the preparation of the manuscript. Valuable suggestions were

FIGURE 1.—*Map of Newfoundland*
Showing location of geographic features mentioned in text.

gratefully received from Dr. A. F. Buddington, Dr. H. H. Hess, and Dr. Kenneth DeP. Watson, of Princeton University.

EARLIER OBSERVATIONS ON FAULTING

In 1866 Alexander Murray, who was investigating the Carboniferous basins in southwestern and west-central Newfoundland, stated (Murray and Howley, 1881, p. 90) that there seemed to be evidence that a "great fault . . . intersects the island diagonally from shore to shore, running in an almost perfectly straight line from near the entrance of the Little

Codroy River to White Bay." This fault has a length of about 175 miles, and if for no other reason than its regional extent would be regarded as a significant structural feature.

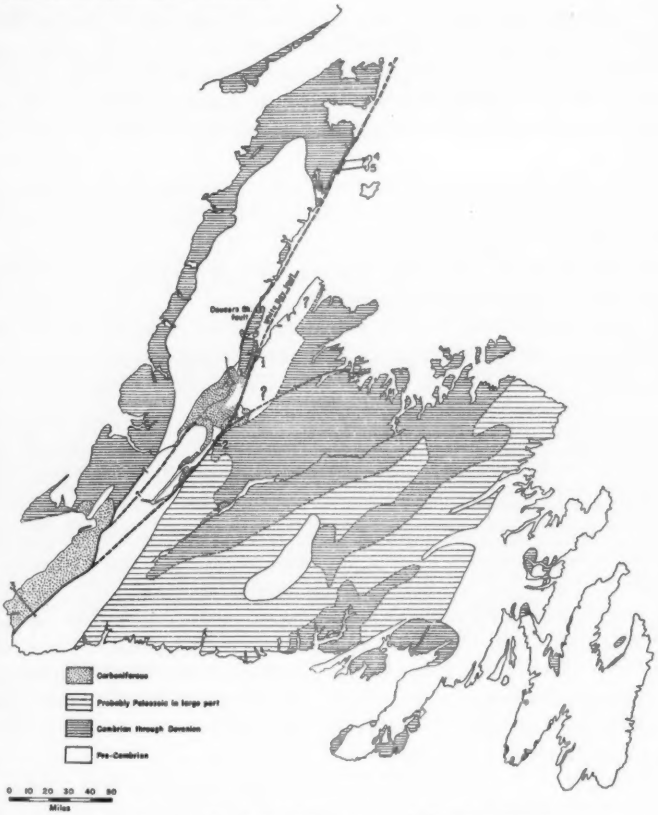

FIGURE 2.—*Geologic sketch map of Newfoundland*

(a) Distribution of pre-Cambrian, Paleozoic (except Carboniferous, and Carboniferous rocks); (b) Fault lines discussed in text; (c) Location of cross sections shown in Figures 3 and 4. Redrawn and modified from a map by A. K. Snelgrove (1938, Pl. II).

The evidence presented by Murray (Murray and Howley, 1881, p. 87-91) was the following:

(1) The Carboniferous on the southwestern coast, in the region from Trainvain Brook (Fig. 1) to the vicinity of Cairn Mountain, was found, "wherever visited," to be strongly deformed. For example, between Grand and Little Codroy rivers (Figs. 1, 2),

"The measures both of the coast and the river sections were found uniformly to dip to the south-eastward . . . toward the Cape Ray range of mountains, until within a very short distance of the gneiss, where they are usually either highly tilted, inclining in the opposite directions, or vertical."

(2) At Deer Lake (Fig. 1) the Carboniferous conglomerates dip northeastward at low angles on the western side of the lake. The same rocks are vertical or dipping westward on the eastern side. The disturbed condition was interpreted as being "evidently connected with a fault running in a north-easterly direction, with a downthrow on the north-west side."

(3) The usually horizontal Carboniferous strata on the Humber River above Deer Lake are "somewhat tilted, with a dip toward the northwest," near Johns Falls (Fig. 1).

(4) The "Silurian" (now Ordovician) limestones at the head of Spruce Brook (Fig. 1) are much disturbed.

(5) The coarse conglomerates on Miller Island (Fig. 1), at the head of White Bay, and on the western shore of that bay are vertical.

In 1873 Murray (Murray and Howley, 1881, p. 331) remarked:

"There seems also to be good ground for suspecting that a bifurcation of this fault takes place near the mountain gorge of the Middle Barachois, which probably communicates with the . . . fault on the southeast side of the Grand Pond (now called Grand Lake), and finally follows the depression of the Indian Brook towards Hall's Bay; maintaining almost a straight course for the whole distance. From the place of the supposed bifurcation the more northerly fault runs . . . at the foot of the mountains (Long Range), keeping a straight course up to . . . Cairn Mountain, and a few miles beyond . . .; a little north of which, near the Little Barachois, a second bifurcation appears to take place, the most westerly fault running on the eastern side of the valley of Harry's Brook, . . . bearing thence towards the lower reach of the Humber River. The eastern branch of the fault is the one alluded to in the Report of 1866, as striking upon the south-east side of Deer Pond, and thence up the valley of the Humber River to White Bay."

Howley, who succeeded Murray as Government Geologist, did not pursue the examination of this feature. Commenting on Murray's postulated line, Howley (Perret, 1913, p. 49) stated that "The fault was merely theoretical and did not in reality exist."

It was not until 1930 that Murray's fault was introduced again. At that time, Snelgrove compiled a structural map of the island and suggested a relationship between a thrust fault in western Notre Dame Bay[2] and the line drawn by Murray from Grand Lake through the valley of Indian Brook.

EVIDENCE FOR ZONE OF WEAKNESS

GENERAL STATEMENT

In recent years geological studies have been made in the Carboniferous belt of the southwestern coast (Bay St. George area), at Grand Lake,

[2] This map is unpublished, but the fault in question is discussed by Snelgrove (1931, p. 508) as the Wild Bight thrust.

and in White Bay. In most of the other parts of the zone of weakness, as originally defined, Murray's observations must still be accepted, insofar as they apply to the attitude of beds and to the lithology of the sediments.

SOUTHWESTERN COAST

Hayes and Johnson (1938, p. 7), who studied the Carboniferous of the Bay St. George area (Fig. 1), which consists of a sequence of nonmarine and marine Mississippian sediments and terrestrial Pennsylvanian rocks, state:

"Igneous or metamorphic rocks of the Long Range, apparently in thrust contact with the Barachois series (Pennsylvanian), are exposed in the valleys of some of the tributaries of the Little Codroy and Grand Codroy Rivers for a distance of 25 miles. The actual fault is also well exposed to the north along Sheep Brook, a northern tributary of Flat Bay Brook, north of Cairn Mountain. It is uncertain whether there is a continuous faulted boundary in the central part of the map area for the Carboniferous sediments are not quite so highly tilted along Crabbs, Middle Barachois, and Robinson Rivers as in the localities where the fault was observed . . ."

It seems safe therefore to assume that there has probably been faulting over a distance of 75 miles. It has been actually observed over somewhat less than half this distance. The structure sections prepared by Hayes and Johnson for this area indicate a high-angle thrust of the pre-Cambrian [4] northwestward over the Carboniferous.

This area carries the fault past the first bifurcation mentioned by Murray at Middle Barachois River and continues to the second bifurcation north of Cairn Mountain (Figs. 1, 2).

The fault marks the boundary between the Long Range and the coastal lowland. The front of the Long Range is a fault-line scarp, an abrupt rise from the lowland to the high plateau surface (Twenhofel and MacClintock, 1940, p. 1675).

Apparently related in time of development to the principal fault are numerous high-angle faults in the Carboniferous coastal belt (Fig. 3). The tilted and folded sediments are broken by faults with various trends, some almost perpendicular to the main fault.

MIDDLE BARACHOIS RIVER TO GRAND LAKE

The eastern branch in the first bifurcation of the fault postulated by Murray proceeds from Middle Barachois River across the Long Range to Grand Lake. This section of the Long Range consists entirely of pre-Cambrian rocks (Fig. 2). There is no known published information regarding faulting in the Long Range. However, Buddington (personal

[4] For the present, the igneous and metamorphic rocks of the Long Range will be classified as pre-Cambrian, following Hayes and Johnson. Recently, Twenhofel and MacClintock (1940, p. 1675) have remarked: ". . . all that is really known is that they are older than the Mississippian. . . . It is probable that some of the intrusives are of Paleozoic age."

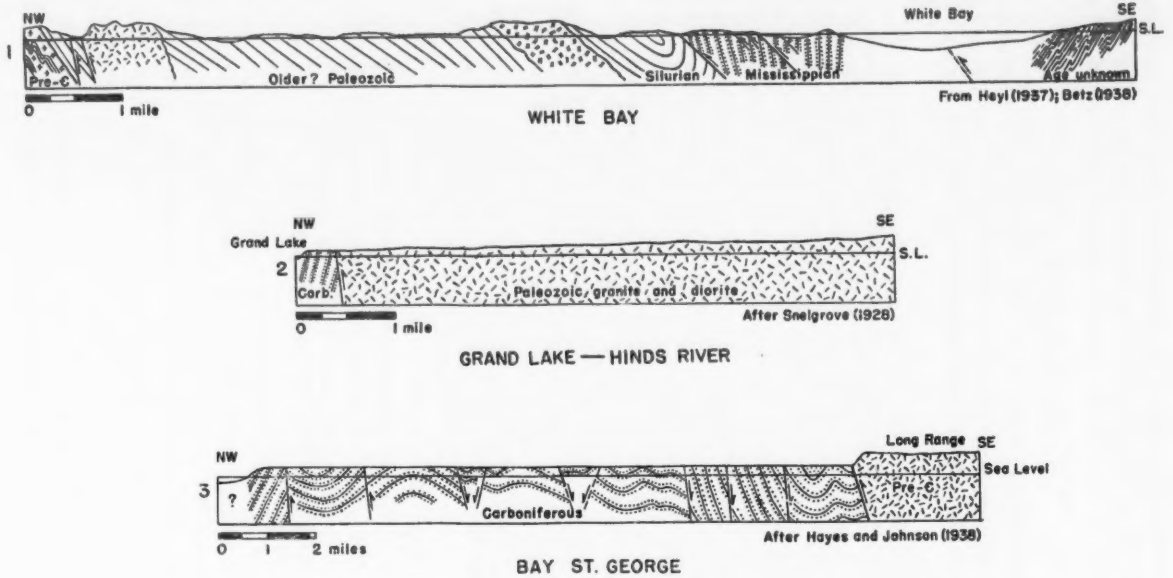

FIGURE 3.—*Structure sections*

Across zone of faulting from southwestern coast to White Bay. *See* Figure 2 for location of sections. Sources indicated under each section. No vertical exaggeration.

communication), who in 1939 made a reconnaissance inland from Bay St. George, says:

"About 15½ miles east of St. Georges in the valley of Flat Bay Brook, just east of the anorthosite mass (Cairn Mountain), there is an exposure of intensely fractured and slickensided granite which might well be the locus of a fault zone."

This deformed granite lies within the zone of weakness projected from the southwestern coast to Grand Lake.

GRAND LAKE

On the eastern side of Grand Lake (Fig. 1), the fault was said by Murray (Murray and Howley, 1881, p. 526) to be

"well pronounced for a great part of the whole length of the lake, where the various members of the Carboniferous series butt up in highly disturbed strata against a mass of greenstone which follows a very straight course along the western flank of the gneissoid mountains, pointing in the direction of the inlet into Sandy Pond (now Lake)."

Snelgrove (1928, p. 1092-1093) examined a section on Hinds River (Fig. 1), a stream flowing into Grand Lake near its northeastern end, and estimated that more than 1000 feet of Carboniferous arkoses rest unconformably on andesite-diabase flows (presumably Paleozoic) at the mouth of the river. In a structure section, Snelgrove indicates a high-angle fault in which Paleozoic igneous rocks are moved northwestward against the Carboniferous (Fig. 3).

To the north of Hinds River, the fault was recognized by Murray (1881, p. 526) on Coal Brook, another stream emptying into the lake.

Pronounced disturbance of the Carboniferous strata on the western side of Grand Lake, apparently due to faulting, was also noted by Murray (1881, p. 63).

LITTLE BARACHOIS RIVER TO DEER LAKE

In 1866, Murray pointed to the disturbed condition of the "Silurian" (now Ordovician) limestones at the head of Spruce Brook (Fig. 1), a stream flowing into Grand Lake. The course of Spruce Brook lies wholly within the pre-Cambrian of the Long Range (Fig. 2). The contact of the Ordovician with the pre-Cambrian is indicated about 4 miles west of the head of this stream. The writer believes that the evidence should be referred more properly to the branch of Murray's fault drawn from Little Barachois River to the lower Humber. No further evidence for the fault from the southwestern coast to Deer Lake seems to have been presented.

DEER LAKE

The disturbance of the Carboniferous beds on either side of the lake was related to faulting by Murray (1881, p. 526), who in 1879 remarked

that the evidences of the great fault crossing the island were "very distinctly displayed . . . at the north-east end of Deer Pond," among other places.

HUMBER VALLEY

Most of the Humber Valley from Deer Lake to the vicinity of White Bay and the area from the northern end of Deer Lake to Grand Lake and Sandy Lake are occupied by Carboniferous rocks (Murray and Howley, 1881, p. 86, 523; 1918, p. 160, 184). In the Humber Valley, Howley (Murray and Howley, 1918, p. 184) described them as "spread out in nearly horizontal strata, or in low, wavelike undulations." However, Murray (1881, p. 523) also reported "the existence of a series of folds and sharp flexures which affect the distribution of the different members of the group." The tilting of the beds at Johns Falls, given in the original discussion of his fault, has already been mentioned.

The attitude of the beds and scarcity of exposures hindered the recognition of the stratigraphic succession. Nevertheless, Murray (1881, p. 523-524) listed a sequence of Carboniferous beds totaling 4080 feet in thickness, admittedly an approximation. Speaking then of faulting in the Humber Valley he (1881, p. 526) says:

> "In my Report for 1866 . . . a great fault is represented, intersecting the island diagonally, the course of which passes a little east from the bed of the west branch of the (Humber) river, running in the direction of White Bay . . . but the nature of the country between the branches of the Humber is such as to prevent the possibility of tracing it there; and the usual flat character of the land, together with the apparently undisturbed state of the strata wherever seen, lead to the inference that probably the dislocation, while contemporaneous with the lower measures, is older than No. 4 (unit of the Carboniferous series; about the middle of the section), and that it, in common with the lower strata, is covered over by the higher measures unconformably."

It is thus apparent that Murray had no strong evidence for drawing his fault through the upper Humber Valley.

WHITE BAY

In the region around the northern end of the fault defined by Murray, recent work has been done by Heyl (1937) and by the writer.[5] Murray found evidence for the fault in the strong tilting of coarse conglomerates on Miller Island and on the western shore of the bay. This observation is quite correct but is not adequate to establish the fault.

The conglomerates belong to a sequence of nonmarine, plant-bearing Mississippian beds. Prior to 1937 these rocks were called Devonian on the basis of plant fossil determinations by Dawson (Murray and Howley, 1881, p. 45). The Mississippian strikes north-northeast in a belt

[5] The results of a study of the southern part of White Bay, based on field work in 1938, are to be published by the Geological Survey of Newfoundland.

bounded on the east by White Bay. Multitudes of small thrusts and shear planes traverse the beds.

On the eastern side of White Bay, two main groups of metamorphic rocks are exposed. In the southern part of the bay, the rocks are mainly quartz-mica schists and crystalline limestones. From the northern end of White Bay and in the area extending eastward to Fleur-de-Lys (Fuller, 1941) and Baie Verte (Watson), somewhat similar rocks have been described. The two groups are unlike any others known in northern or western Newfoundland but may possibly be related to pre-Cambrian schists of the Northern Peninsula, or they may be the metamorphosed equivalents of Paleozoic sediments.

Since the metamorphics in White Bay dip uniformly westward and the Paleozoic rocks dip eastward, Howley (Murray and Howley, 1918, p. 495) considered the bay to be a large synclinal trough. He thought it possible to correlate different units of the metamorphic group with parts of the Paleozoic section. Murray (1881, p. 18) had held this view in 1864 but appears to have discarded it when discussing the transinsular fault.

Present knowledge of the Paleozoic and pre-Cambrian in this region tends to vitiate the correlation of the rocks on opposite sides of the bay. Therefore, the synclinal interpretation of White Bay has become doubtful. There is no apparent reason why the unmetamorphosed Mississippian sediments on the western shore, though highly deformed, should find counterparts in a series of schists on the eastern side of the bay, which come within a few hundred yards of each other. The discordance between the sections strongly suggests major dislocation along the bay (Fig. 3).

The bay itself is the principal topographic feature of the region. White Bay is the drowned northern extremity of a basin that extends southward to Grand Lake. The lowland is sufficiently pronounced to be recognized as a separate physiographic division of the island (Twenhofel and MacClintock, 1940, p. 1699-1701). The few outcrops seen in the lowland are sheared and deformed similarly to those on the western side of White Bay. There seems to be a strong suggestion of a fault zone throughout the lowland and of a direct relation between the lowland and the faulting. The abundance of north-northeast-trending faults, not only in the Mississippian but also in the older Paleozoic rocks, points to extensive regional disturbance in White Bay.

Heyl, in studying the Sops Arm area of White Bay, found a prominent fault line (Doucers Brook fault) (Fig. 2) at the foot of the Long Range Mountains, between the pre-Cambrian gneiss and the Paleozoics.

The fault, where observed, indicates a high-angle thrust to the northwest. Heyl (1937, p. 19) suggests that the fault at Deer Lake "may possibly be the southward continuation of the Doucers Brook fault." The writer prefers to refrain from connecting the Doucers Brook and Deer Lake faults until more is known about the Humber region.

SUMMARY

The preceding discussion of evidence for faulting may be summarized as follows:

(1) The zone of faulting has been confirmed by observation on the southwestern coast, as far north as Little Barachois River.

(2) Owing to lack of observation, a fault zone from Little Barachois River to Deer Lake and from there to White Bay, proposed by Murray, cannot be established.

(3) On the line connecting Middle Barachois River and Grand Lake, intense fracturing has been noted, which may point to the presence of a fault zone.

(4) Faulting was reported to be pronounced on the eastern side of Grand Lake by Murray. It has been confirmed at one place near the northern end of the lake.

(5) A fault in the White Bay basin is strongly suggested by the extensive thrust faulting in the Paleozoics paralleling the western side of the bay, by the dissimilarity of the sections on the opposite sides of the bay, and by topography.

(6) Grand Lake and White Bay are connected by a lowland in which conditions similar to those in White Bay prevail.

Some revision of Murray's fault lines seems to be necessitated by these findings. The principal fault zone crossing the island is therefore projected from the southwestern coast across the Long Range to Grand Lake and White Bay.

Connection of the Deer Lake fault with the Doucers Brook fault in the White Bay region was proposed by Heyl (1937). The connection is not made here because of the uncertain evidence of faulting in the Humber valley between Doucers Brook and Deer Lake.

POSSIBLY ASSOCIATED FAULTS

GENERAL STATEMENT

In the preceding discussion, certain sections of the original lines drawn by Murray have been omitted. Brief comment follows on these sections and on faults that may be possibly associated with the main zone.

GRAND LAKE TO HALLS BAY

An important high-angle thrust extends for about 25 miles approximately east-west along the southern margin of Notre Dame Bay (Espenshade, 1937, p. 20). It turns southwestward in Halls Bay and has been traced into the valley of Indian Brook (MacLean, unpublished). Apparently the thrust fault in Halls Bay and Indian Brook valley corresponds to the northern end of the line drawn by Murray from Grand Lake to Halls Bay.

Another possible connection of this line was suggested by Snelgrove, as previously noted. According to this proposal, a north-northeast-trending thrust fault at Wild Bight (Fig. 1) on the western shore of Notre Dame Bay might be regarded as a continuation of Murray's line from the valley of Indian Brook. Recently, a number of faults with more or less the same trend were described (MacLean) in the area northwest of Halls Bay. These constitute a fault zone, which seems to converge southwestward toward Indian Brook.

From this it appears possible that Murray's fault may branch near its northern end. Unfortunately, no evidence has been presented since Murray's time for drawing the line from Grand Lake into Indian Brook valley. This connecting link, if proved, would relate faulting in the southwestern coast-Grand Lake-White Bay line to that in Notre Dame Bay.

LITTLE BARACHOIS RIVER TO LOWER HUMBER RIVER

The fault in this section was located by Murray on the eastern side of Harrys Brook (Fig. 1). The position of the fault along this line was defined by only one point. It was said (Murray, 1881, p. 331) to be "observable in the Silurian (now Ordovician) strata around Es-que-dege-weh Gospen (Fig. 1), at the head of the eastern branch" of Harrys Brook. Presumably it parallels, or coincides with, the Paleozoic-pre-Cambrian contact.

Recent studies in the Bay of Islands area (Fig. 1) have not reached as far eastward as the postulated fault. The importance of faulting throughout the coastal belt between Bay St. George and Bay of Islands, and northward to Bonne Bay (Fig. 1), is recognized. Ultramafic intrusive bodies form prominent features in this part of the coast and are involved in faulting that has been interpreted as overthrusting (Cooper, 1936; Buddington and Hess, 1937). The associated Ordovician sediments, particularly those between the intrusives and the pre-Cambrian highland, show extreme deformation. It is reasonable to assume that the pre-Cambrian block has also been affected by the

compressive forces and that a fault will be found at the contact with the Paleozoics.

WESTERN COAST NORTH OF BONNE BAY

Johnson (1941b, p. 144) finds similar relationships in the region north of Bonne Bay. From the Highland of St. John (Fig. 1) southward, low-angle and high-angle thrust faults have produced intense disturbance in the Cambrian and Ordovician. As in the Bay St. George area, the eastern border of the coastal belt is marked by a high-angle fault. Over much of the northwestern coast, it is marked by "The great Long Range Thrust, bringing up the Pre-Cambrian crystallines over the Paleozoic foreland . . ." Sections showing the thrust fault at two points on this coast have been published by Johnson (1941a, p. 1560-1561).

Faulting in this section of the coast apparently occurred more than once during the Paleozoic. Johnson (1941b, p. 145) states that:

> "There is no accurate means of dating the periods of faulting in this area, but faults corresponding to the high-angle thrusts appear to be post-Carboniferous further south. It is suggested that the earlier period of low-angle thrusting may well be Taconic or Caledonian in age, while the high-angle faults are Appalachian in age."

A connection of the fault line on the southwestern coast with the Long Range thrust to the north cannot be established at present. However, the same fault relation between pre-Cambrian and Paleozoic rocks is seen wherever this boundary has been examined from the southwestern tip of the island to the Highlands of St. John. It is significant that Appalachian faulting is emerging as an ultimate cause for many of the present stratigraphic, structural, and topographic discontinuities in western Newfoundland.

NATURE OF FAULTING

In discussing the evidences for faulting, some mention was made of the nature of the faults. From these remarks it is seen that there are few direct observations along the principal zone of weakness. The lack of data in the areas that have been studied is due probably in large part to poor exposures or lack of exposures on the line of the main dislocation.

Hayes and Johnson (1938, p. 7) have observed the fault at the contact between the pre-Cambrian and the Carboniferous in the Bay St. George area. Their structure sections (Hayes and Johnson, 1938, Pl. I) indicate the dip of the fault as about 65°-75° E., with relative movement of the pre-Cambrian northwestward over the Carboniferous. An earlier note by Mook (1926, p. 349) regarding this fault at South Branch (Fig. 1) on Grand Codroy River includes a section in which the dip is shown at about 40° E.

In the White Bay region, many small thrust faults are observed in the Mississippian on the western shore of the bay, which are believed to be related to the fault that has determined the position of the bay itself. There is a wide variation in dips on the fault planes. Slumping has possibly disturbed the original attitude in some cases. The strike is regularly north-northeast. Near the southern entrance to Sops Arm, a major thrust fault of the region shows a dip of approximately 45° E. Drag indicates overthrusting to the northwest (Heyl, 1937, p. 20, Fig. 3). This seems to be typical of the faulting in the Mississippian of White Bay.

Elsewhere in the principal line of dislocation, the faults are located mainly by indirect evidence. The north-northeast trend of the structure is maintained with a suggested compression in a southeast-northwest direction. Normal faults, which Murray always visualized in discussing this zone, may be present locally. In the southwestern coastal area, blocks of deformed Carboniferous sediments are downdropped along high-angle faults striking obliquely to the main thrust fault, "which may well be interpreted as the result of shortening," according to Hayes and Johnson (1938, p. 8).

NORTHWARD PROLONGATION OF ZONE OF WEAKNESS

The eastern coast of the Northern Peninsula from the head of White Bay northward is remarkably straight (Fig. 1) suggesting that the coast is defined by a fault line. This assumption is strengthened by the presence of a zone of faulting in White Bay that parallels the coast.

Thrust faults with north-northeast trends are the principal structural features on this coast. This has been observed in the Paleozoic (Cambrian and Ordovician) rocks in Canada Bay (Betz, 1939, p. 31), in Hare Bay (Cooper, 1937, p. 13), and at Great Harbour Deep (Fig. 1) (Foley, 1937, p. 14-16), where the faulting affects pre-Cambrian rocks but may be Paleozoic.

The only places where a fault located immediately on the coast would come into view are the Cape Rouge and Cape Fox peninsulas (Fig. 1) between Canada Bay and Hare Bay (Fig. 4). The small peninsulas are composed of Mississippian conglomerates, sandstones, and siltstones, which are thrust against the Ordovician rocks exposed on the coast from Canada Bay northward.[6] These peninsulas are of further interest because they contain the most northerly exposures of Mississippian sediments in Newfoundland and thereby afford evidence of the time of deformation. The Mississippian is also found

[6] Personal communication, Dr. Helgi Johnson.

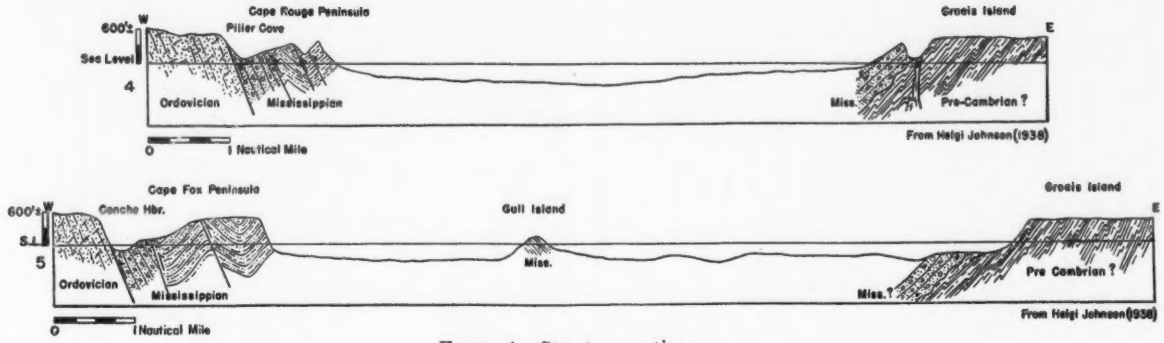

FIGURE 4.—*Structure sections*

Cape Rouge and Cape Fox peninsulas to Groais Island. After Helgi Johnson.

offshore on Groais Island (Fig. 1), where it occupies the northwestern tip of the island, lying unconformably upon schists of unknown age.

In Hare Bay, low-angle thrusting has been extremely important. A large *decke* has been carried more than 11 miles west by the Hare Bay thrust over already folded and thrust-faulted Ordovician sediments. The front of the thrust sheet is traced from the northern shore of Hare Bay to Pistolet Bay (Fig. 1). The thrust thus involves the whole northeastern tip of the island. The thrust plane dips eastward at a very low angle, passing gradually beneath Hare Bay. The sheet is truncated by the coast. Its dip seaward is a matter of conjecture.

In considering the northward prolongation of the zone of weakness from White Bay, the particular line chosen has been the Doucers Brook fault, following an earlier suggestion by Heyl (1937, p. 19). The Doucers Brook fault lies immediately adjacent to the coast when extended in its trend from Coney Arm (Fig. 2). The White Bay fault, if projected in a straight line, would lie several miles offshore, between Groais Island and Cape Rouge Peninsula. Unfortunately there is no place along the coast where it is possible to identify two principal faults, such as the White Bay and Doucers Brook faults, nor can the main thrust in the Cape Rouge-Cape Fox area be determined as the northward extension of one of these faults.

AGE OF FAULTING

The principal zone of weakness, as it is now outlined, involves the two large belts of Carboniferous rocks in Newfoundland. The faulting, whether seen or postulated, must be late Paleozoic or younger.

On the southwestern coast the earliest possible age for the thrusting is post-Early Pennsylvanian. The youngest rocks in the thrust zone are dated "Lower or Middle Pottsville age or Westphalian A" (Hayes and Johnson, 1938, p. 22). The compressive movements which produced the deformation of these Carboniferous sediments are referred to the Appalachian Revolution.

In the Carboniferous belt on the eastern side of the Long Range the rocks have not been dated so precisely. In the Humber Valley and at Grand Lake, Mississippian and Pennsylvanian rocks are reported. At Deer Lake, the beds are apparently Early Mississippian. The plant fossils in the beds in White Bay and on Cape Rouge and Cape Fox peninsulas and Groais Island indicate an Early Mississippian age. Thus, at Deer Lake and in the region from Grand Lake northward, the faulting can be determined only as post-Early Mississippian.

Whether all the faulting (probably thrust faulting in all cases) was

contemporaneous or not, it can be classed with the late Paleozoic deformation of eastern North America. As such it demonstrates the presence of the Appalachian Revolution in Newfoundland.

Following the Paleozoic, Newfoundland appears to have been an emergent area. No record of sedimentation is known from Pennsylvanian time until the Pleistocene glacial deposits were formed. The latest fault movements can therefore not be assigned to a post-Paleozoic time.

Mention has been made of faults associated with the main zone of weakness but whose age is not necessarily the same. The Doucers Brook fault involves none of the Mississippian rocks in the White Bay region. It has been dated by Heyl as probably Caledonian. As stated before, Heyl (1937, p. 21) considers renewed movement probable along this fault in post-Mississippian time. If the extension of the Doucers Brook fault along the coast to Cape Rouge Peninsula be accepted, it must follow that there was movement along this line during the late Paleozoic.

The faulting of Paleozoic rocks in Hare Bay and Canada Bay is, of necessity, post-Ordovician (or post-Middle Ordovician), since the youngest rocks affected are of that age. Now that post-Mississippian thrust faulting is known in the Cape Rouge-Cape Fox area, between these bays, late Paleozoic deformation or at least renewed movement on older fault lines must be considered for the entire Paleozoic belt.

SUMMARY

(1) A fault zone extending from the southwestern coast of Newfoundland in a northeasterly direction to Grand Lake and White Bay is indicated by the available evidence.

(2) In each area along this zone, on which there is information, direct or indirect evidence points to high-angle thrusting. The fault planes dip eastward, and thrusting is considered to have been to the west.

(3) The faults on the southwestern coast and in the region from Grand Lake to White Bay involve Carboniferous rocks. It is therefore probable that the faulting can be assigned to the late Paleozoic and that it is related to the Appalachian Revolution.

(4) The eastern coast of the Northern Peninsula has been thought to be defined by a fault line. Evidence of the late Paleozoic thrust faulting is found at Cape Rouge and Cape Fox Peninsulas. Other faults in the coastal region of the Northern Peninsula and in White Bay, hitherto poorly dated, may be in part late Paleozoic or, if older, may have experienced renewed movement during the Appalachian orogeny.

(5) A high-angle thrust fault, in which the pre-Cambrian moved over Paleozoic rocks, on the northwestern coast of the island, has recently been suggested to be Appalachian. It is possible that a connection of this fault with the late Paleozoic thrust fault to the south will be established.

(6) Thrust faults in western and southern Notre Dame Bay lie at the northern end of a fault line that was postulated in 1873 as extending northeastward from Grand Lake. No evidence has been found to support this fault line.

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PHOSPHATIZATION AT MALPELO ISLAND, COLOMBIA

BY DUNCAN McCONNELL

CONTENTS

	Page
Abstract	707
Introduction	707
Acknowledgments	709
Petrography and mineralogy	709
Augite andesite	709
Amygdaloid	710
Phosphate rock	712
Discussion and summary	713
References cited	714

ILLUSTRATIONS

Plate	Facing page
1. Malpelo Island	708
2. Photomicrographs	709

ABSTRACT

A petrographic study is presented of three specimens from Malpelo, a volcanic island in the eastern equatorial Pacific Ocean. Two of the specimens show appreciable alteration by phosphatic solutions from guano: an amygdaloid in which the feldspars have been completely destroyed and replaced and a phosphate rock composed essentially of phosphosiderite and strengite, a dimorphous pair with the composition $FePO_4 \cdot 2H_2O$. The third is a comparatively unaltered augite andesite having a rather low silica content and possibly containing anemousite. The petrologic relations among aluminum-iron and calcium phosphates are discussed briefly.

INTRODUCTION

During the voyage of the schooner ASKOY (Murphy, 1941), Dr. Robert Cushman Murphy and one other member of his party landed at Malpelo Island in March 1941. On this occasion a number of samples of Malpelo rocks were assembled, but the roughness of the sea would not permit their removal from the island. The specimens accessible for the present study were obtained by Dr. R. C. Murphy from Comandante Eduardo Fallon of the Colombian Navy, who was a member of the scientific staff of the schooner ASKOY. Comandante Fallon collected them on December 22, 1940, when he visited Malpelo Island in the gunboat JUNIN, in company with Mr. G. William Bylander, Swedish Consul General at Buenaven-

tura. These rock specimens consist of an augite andesite, an amygdaloid, and a phospate rock high in iron.

Malpelo Island, a territory of the Republic of Colombia, lies 250 nautical miles west of Buenaventura, Colombia, at Lat. 3° 59′ N., Long. 81° 34′ W. It is a volcanic island approximately 2 kilometers long and ½ kilometer wide, but having a maximum elevation of 258 meters. It has been described as resembling a large black iceberg (Pl. 1, figs. 1, 2) with clusters of rocks northward and southward. Very deep soundings only a short distance from its cliffs indicate that the island rises very steeply from appreciable depths. No beaches and few ledges are present, and landing is possible only at one or two places and then only when the ocean is calm.

Although Malpelo Island was known as early as 1530, landings have seldom been made, and scientific expeditions rarely visit the island. It is virtually devoid of vegetation except for algae in the spray zone, although grass and sparse shrubbery have been observed on one or more of the outlying stacks. The vertebrate fauna is said to consist of two species of lizards and various sea birds. Of the avifauna, about a dozen species are known, but the only one present in significant numbers is a large, white booby *(Sula dactylatra)*. However, even these are sparsely represented when compared with the bird populations of the Peruvian guano islands—perhaps 15,000 pairs as compared with several million for a Peruvian island of comparable size.

There is no significant accumulation of guano on the island at present, and, if the topographic configuration has always been essentially similar, there probably never have been guano deposits of any consequence. The phosphatization of the lavas, then, is probably limited to portions adjacent to numerous fractures where the organic phosphate solutions could accumulate. The rocks are extensively fractured, and this condition has resulted in the development of no less than 11 sea caves and several arches. Water is said to ooze from the rocks 20 to 50 feet above the sea throughout the entire year.

Deposits of volcanic ash are present, and, if sufficiently porous, these might act as reservoirs for the phosphatic solutions. According to Murphy (personal communication), Malpelo seems to be composed of dense volcanic igneous rocks interlayered with scoria and ash. Much of the surface is extraordinarily weathered and clinkery. Vesicles filled with silica are conspicuous in the lava, and siliceous replacement bodies occur in some of the ash.

Regardless of any limitations in extent of the deposits, the phosphatization has resulted in almost complete replacement of the primary minerals of a basic to intermediate volcanic rock. Significant alteration of the

FIGURE 1. VIEW FROM SOUTH
Southwest tip of island and two-thirds of southeast coast.

FIGURE 2. VIEW FROM NORTH
Part of northwest coast and outlying stacks.

FIGURE 3. ABOVE LANDING COVE
Fractured nature of volcanic rocks and masked booby (*Sula dactylatra*).

FIGURE 1. AUGITE ANDESITE
Fluxional arrangement and inclusions in a feldspar phenocryst. ×26

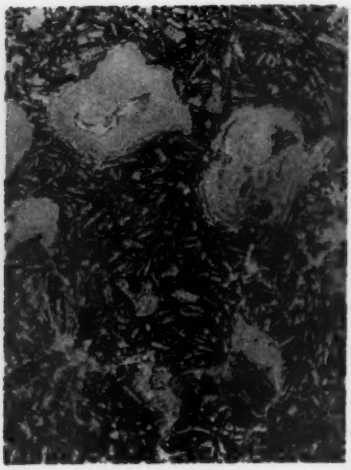

FIGURE 2. AMYGDALOID
Irregular amygdules of aluminum phosphate and relict feldspar laths. ×33

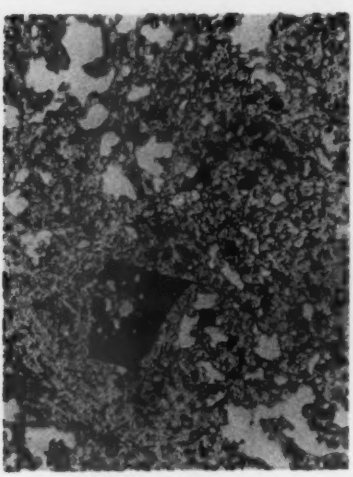

FIGURE 3. PHOSPHATE ROCK
Relict phenocryst of titaniferous magnetite and fluxional arrangement of vesicles. ×33

FIGURE 4. PHOSPHATE FROM GRAN ROQUE
Spherulites of barrandite associated with cloudy collophane. ×21

PHOTOMICROGRAPHS

primary minerals of the amygdaloid has taken place also. In addition, the low silica content of the augite andesite cannot be reconciled with its mineralogical composition.

ACKNOWLEDGMENTS

The petrographic investigation of the Malpelo rocks was undertaken at the suggestion of Dr. Frederick H. Pough, of the American Museum of Natural History. All the general statements concerning the island are based upon publications (Murphy, 1936; 1941) and personal communications from Dr. Murphy, who also supplied the photographs of the island. Mr. William Y. Holland prepared the photomicrographs.

Mr. Howard M. Hunter chemically analyzed the phosphate rock and prepared the spectrographic plates used for qualitative determinations. Messrs. John L. Gilliland and Dale Torgeson furnished valuable advice concerning the analytical procedures. Mr. Lisle L. Wheeler, of the University of Texas, prepared the X-ray diffraction photographs.

The specimen from the island Gran Roque, D. F., Venezuela, which is compared with specimens from Malpelo, was furnished by Mr. S. E. Aguerrevere, of the Servicio Técnico de Minería y Geología, Caracas. Specimens of aluminum phosphate from the Connétable Islands, French Guiana, were furnished by the U. S. Department of Agriculture, through the kindness of Dr. K. D. Jacob.

The manuscript and the thin sections of the rocks were examined by Professor Frank F. Grout, whose comments have been most helpful.

PETROGRAPHY AND MINERALOGY

AUGITE ANDESITE

This dense aphanitic rock is superficially weathered to depths of 1 to 5 mm. Weathered portions are medium brownish gray, whereas the interior is dark steel gray.

Thin sections appear virtually holocrystalline, microporphyritic, and diabasic, with distinct fluxional arrangement (Pl. 2, fig. 1). The plagioclase laths show continuous gradations in composition, being more calcic toward their interiors. They range from intermediate andesine to basic oligoclase. Potash feldspars were not observed, nor were any colorless constituents with indices lower than that of Canada balsam. The mineralogical composition was estimated as follows:

	Per cent by volume		Per cent by volume
Plagioclase	50	Apatite	< 1
Augite	25	Leucoxene	5
Magnetite (titaniferous?)	10	Chloritic minerals	8
		Quartz	2

The small amount of quartz is presumed to be secondary. A euhedral hexagonal section observed in a matrix of chlorite gave an interference figure with positive sign, eliminating the possibility of nepheline. All the grains identified as quartz exhibited slightly greater clarity than the feldspars.

The occasional phenocrysts of pyroxene appear almost unaltered. They have $Z \wedge c \approx 44°$, and $Z_v \wedge c > Z_\rho \wedge c$ and therefore probably correspond to diopsidic augite, which is the pyroxene observed as phenocrysts by Barth (1931) for a number of Pacific lavas. The pyroxene of the mesostasis is so finely granular as to elude the determination of its properties. Thus, because of the rather slight differences in refringence, it is impossible to decide whether or not olivine is present in the groundmass. However, phenocrysts of olivine were not observed.

A magnetic ore mineral (probably titaniferous magnetite) occurs in appreciable amounts, but apatite is comparatively rare. Although chloritic minerals are prominent, only very small amounts of finely granular carbonates are present, and epidote was not detected. Leucoxene was estimated at 5 per cent.

As a basis for comparison with the phosphate rock, three chemical constituents of the andesite were determined by Mr. Hunter, as follows:

SiO_2 48.63 TiO_2 3.12 MnO 0.18 (weight per cent)

Although the rock has been tentatively classified as an augite andesite on the basis of the composition of the feldspars and ferromagnesian mineral, this silica content is surprisingly low. Thus, the possible occurrence of anemousite must be considered, particularly since Barth (1930) believes this feldspar is present in the basaltic rock variety, pacificite, which may be related. Without more complete chemical data, it is impossible to decide this matter; inasmuch as a more intensive study of the chemistry of this rock has little, if any, bearing on the phosphatization, its composition was not further investigated.

AMYGDALOID

Weathered surfaces of the amygdaloid are very dark purplish and very rugose, apparently as the result of differential solubility. Fresh surfaces range from dark purplish gray to medium brownish gray and exhibit numerous small (1 to 2 mm.) cream-colored amygdules.

The mineralogy of this rock is not readily decipherable with the microscope. All feldspar laths (Pl. 2, fig. 2) have been rather completely replaced, and no twinning lamellae are visible. Some laths now consist of several small, clear, anhedral crystals of quartz, while others are microcryptocrystalline and seem to be the same phosphatic substance that

constitutes some of the amygdules. However, a number of factors preclude the possibility of definite microscopic determination of the latter. Still other relict laths are composed essentially of a chloritic mineral or mixtures of this mineral with quartz or phosphate.

These same substances (secondary quartz, chlorite, and a phosphatic substance) together with considerable amounts of opal are present in the amygdules. Quartz frequently occurs at the margins of amygdules, forming a shell of cloudy anhedral crystals. The other minerals may occur individually, but usually they occur as mixtures. In many amygdules the opal and phosphatic substances alternate as thin concentric shells. The phosphatic material varies greatly (always with high refringence), some of it appearing almost isotropic and some showing low birefringence, indicating its distinct microcryptocrystallinity.

Two samples of the material from amygdules which showed the greatest differences in physical appearance (excluding the chloritic ones) were selected for study by X-ray diffraction methods. The powder patterns of these proved to be essentially identical and indicated a mixture of two isodimorphous minerals, variscite and metavariscite (McConnell, 1940). The composition of these, $AlPO_4 \cdot 2H_2O$, was confirmed by qualitative microchemical tests on the larger sample. The dimensions of the unit cells suggest the presence of a very small amount of iron isomorphously replacing some of the aluminum.

It was originally expected that one of the samples would prove to be a member of the apatite group, particularly in view of the fact that apatite previously had been found as a constituent of a phosphatized gabbro from Gran Roque, D. F., Venezuela (McConnell, 1941). Members of the apatite group were not recognized in the amygdaloid, but their similarities in appearance to some of the observed microcryptocrystalline products, as well as their similarities in modes of occurrence, prevent denial of their possible presence. Likewise, the possible presence of several other phosphate minerals cannot be eliminated; according to C. Frondel (personal communication) whitlockite has been found in phosphate rocks of the West Indies. The X-ray patterns of the material from the amygdules were examined for diffraction lines in addition to those of variscite and metavariscite, but no other identifiable lines were detected.

The mesostasis of the unaltered amygdaloid probably was a dark-colored glass, nearly opaque because of finely disseminated iron oxides. At present it is cloudy and contains nothing identifiable in addition to chlorite, leucoxene, and hydrated iron oxides. Nothing suggests the character of an original ferromagnesian constituent.

Thus the amygdaloid represents a scoria which has been hydrothermally altered and subsequently phosphatized and in which the vesicles have been filled with several minerals. The banding of opal and phosphatic mineral suggests that these minerals formed more or less concurrently, probably at low temperatures.

PHOSPHATE ROCK

The specimen of iron-phosphate rock is quite smooth, dense, and light brownish gray. Broken surfaces showed the presence of numerous small voids so distributed as to suggest flow banding (Pl. 2, fig. 3). The porosity was estimated to be 14 per cent from thin-section examination.

The texture is characterized by very small anhedral crystals, commonly occurring in subspherulitic arrangement but partly in random orientation. X-ray diffraction data indicate the presence of phosphosiderite and strengite, both of which have the composition $FePO_4 \cdot 2H_2O$. A chemical analysis (Table 1) supports the X-ray determination of this dimorphous pair and indicates the presence of a small amount of aluminum isomorphously replacing some of the iron, as had been previously concluded from the measurement of the diffraction lines.

TABLE 1.—*Chemical analysis of phosphate rock from Malpelo Island, Colombia**

(H. M. Hunter, Analyst)

	Weight per cent		Weight per cent
P_2O_5	33.74	F	none
Fe_2O_3	33.40	CO_2	none
Al_2O_3	4.11	$H_2O(+)$	14.90
SiO_2	4.31	$H_2O(-)$	3.80
TiO_2	5.42		
MnO	0.08		99.76

*Inasmuch as the analysis of an iron phosphate is one involving some difficulty, it seems advisable to outline the procedures employed. Phosphoric anhydride and alumina were determined according to the recommendations of Hoffman and Lundell (1933).

Total iron was determined as Fe_2O_3 by the method of Hoffman and Lundell except that the titration was made with dichromate solution using barium diphenylamine sulfonate as an indicator instead of titrating with permanganate. An excellent check was obtained gravimetrically through precipitation of iron as the sulfide and reprecipitation as the hydroxide.

Fluorine was not found although employing the method of Hagen (1934), and its absence permits simpler procedure for silica. The sample used for the silica determination was fused in pyrosulfate, taken up in dilute sulfuric acid, and evaporated to fumes of SO_3. It was cooled, taken up in water, and filtered; the precipitate was then ignited and treated with hydrofluoric acid in the usual manner.

The residue from the silica determination was fused with pyrosulfate, taken up in sulfuric acid, and added to the filtrate from the silica determination. After addition of tartrate to the dilute sulfate solution, iron was removed by reduction with hydrogen sulfide and precipitation in ammoniacal solution. Titanium was then precipitated in acid solution with cupferron and weighed as the oxide. A colorimetric check on the re-fused oxide indicated that the titanium precipitate was not contaminated. Manganese oxide was determined colorimetrically through the use of a spectrophotometer.

Spectroscopic tests indicated that a number of additional elements were either absent or not

If the chemical constituents are calculated as empirical minerals, the results are as follows:

	Weight per cent
Phosphosiderite + strengite	87.2
Ilmenite	3.0
Quartz (including SiO_2 of opal)	4.3
Leucoxene (calculated as TiO_2)	3.9
Moisture, water in opal, and organic matter	1.6
	100.0

The composition of the leucoxene observed in the thin sections is unknown, but in the absence of calcium (Table 1), it is calculated as titanium dioxide. Also, the ore mineral is probably not ilmenite but titaniferous magnetite.

Preliminary examination of thin sections suggested that the ore mineral and leucoxene were present in the phosphate rock in amounts comparable with their volumes in the augite andesite—10 and 5 per cent, respectively. However, it was not possible to secure a reliable estimate of the leucoxene in the phosphate rock by microscopic methods because of the high refringence of the phosphatic minerals. Estimation of the ore mineral by Rosiwal's method gave 5.6 per cent of the volume of the entire rock (including voids), or about 10 per cent, by weight. The latter value does not agree with the result calculated from the analysis, and no explanation of the discrepancy is forthcoming unless the ore mineral is appreciably altered without showing definite evidence of alteration.

Small amounts of quartz and opal were observed in thin sections. However, their relative ages, compared with the phosphates, were not determined. Occasionally the phosphate minerals seemed to show relicts of feldspar laths. The size range and the distribution of the ore mineral were similar to the augite andesite.

Thus the phosphate rock probably represents rather complete replacement of a diabasic rock—probably one similar to the augite andesite. The possibility of its resulting from phosphatization of a tuff cannot be disregarded but seems less probable.

DISCUSSION AND SUMMARY

Although the attack of phosphatic solutions from guano on basic to intermediate igneous rocks commonly results in complex mineral assem-

more abundant than traces: Mg, Cr, Cu, Ba, Sr, V, and Bi. No precipitate of calcium oxalate was obtained by using the method of Hoffman and Lundell.

Water above 105° C. was determined as loss on ignition, fluorine and carbon dioxide being absent. However, this determination may include a very small amount of organic matter.

Part of the water below 105° C. is considered as water of hydration because of the slight deficiency in H_2O (+) necessary to satisfy the formula, $(Fe, Al)PO_4 \cdot 2H_2O$. This is reasonable because Schaller (1912) found a loss in weight for $AlPO_4 \cdot 2H_2O$ of 5.09 per cent at 110° during the first day.

blages, the minerals of the more highly phosphatized rock from Malpelo are confined essentially to a dimorphous pair of ferric phosphates with the composition $FePO_4 \cdot 2H_2O$. The corresponding aluminum-bearing members of the isodimorphous series usually form under these environmental conditions, but the formation of the iron-bearing members under these circumstances has not been previously reported. It is not unusual, however, to encounter these dimorphous minerals occurring together; this seems to be common throughout most of the compositional range of this isodimorphous series.

The mineralogy of the amygdaloid is more complex. This rock also has been phosphatized significantly, but the relict textures and the intimate admixtures of secondary minerals prevent a precise analysis of the phosphatic constituents. The feldspars have been completely destroyed, apparently without the formation of calcium phosphates, such as apatites and whitlockite (Frondel, 1941). Although aluminum phosphates are probably less soluble under these environmental conditions than are the calcium phosphates, the latter sometimes form as intermediate products and coexist with the aluminum phosphates (Pl. 2, fig. 4).

In general, rather complete solution of silica commonly accompanies phosphatization. The destruction of feldspars and other silicates of basic to intermediate igneous rocks apparently is accomplished by the same solutions which form the aluminum, iron, and in some cases, calcium phosphates. The silica thus dispersed may be carried away in solution or may be precipitated as opal, quartz, or chalcedony. Whether or not opal, once having been precipitated, can be redissolved by continued exposure to the phosphatic solutions is uncertain. However, quartz and certain ore minerals (magnetite, chromite, and possibly others) do not seem to be attacked readily by these solutions.

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